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Ice-Ocean Exchange Processes in the Jovian and Saturnian Satellites

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1 **Ice-ocean exchange processes in outer solar system**
2 **satellites**

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Abstract A growing number of satellites in the outer solar system likely have global oceans beneath their outer icy shells. While the presence of liquid water makes these ocean worlds compelling astrobiological targets, the exchange of heat and materials from the deep interior to the surface also plays a critical role in promoting habitable environments. In this chapter, we combine geophysical, geochemical, and geological observations of the Jovian satellites Europa, Ganymede, and Callisto as well as the Saturnian satellites Enceladus and Titan to summarize our current state of understanding of their interiors and surface exchange processes. Potential mechanisms for driving exchange processes upward from the ocean floor and downward from the satellite surface are then reviewed, which are primarily based on numerical models of ice shell and ocean dynamics and complemented by terrestrial analog studies. Future missions to explore these exo-oceans will further revolutionize our understanding of ice-ocean exchange processes and their implications for the habitability of these worlds.

Keywords Ice-Ocean Exchange · Europa · Ganymede · Callisto · Enceladus · Titan

In this chapter, we review ice-ocean exchange processes in outer solar system satellites that are the best candidates to host subsurface oceans: the icy Galilean satellites Europa, Ganymede, and Callisto and the Saturnian satellites Enceladus and Titan. Section 1 summarizes our current state of knowledge of the interiors of these moons, Section 2 describes surface exchange processes, Section 3 describes ice shell dynamics and exchange processes, Section 4 describes ocean dynamics and exchange processes, Section 5 describes terrestrial analogs, and Section 6 concludes with implications for habitability and future exploration.

1 Interiors of Icy Ocean Worlds

Most of what we know about the interiors of known icy ocean worlds comes from the *Galileo* (1989-2003) and *Cassini-Huygens* (1997-2017) missions. As reviewed by Hussmann et al (2015) among others, the interiors of icy satellites are explored through the following data: radius and mass, gravity field, magnetic field, rotational state and shape/topography, surface temperatures and heat flow, composition of surface and atmosphere, activity at the surface, and knowledge of its formation and evolution including surface geology and tectonics, orbital dynamics, and chemical environment during accretion. Complementary to these observational data are laboratory and numerical data on the material properties of water/ice, rock, and metal, as well as their equations of state (e.g., Choukroun and Grasset, 2010; Vance and Brown, 2013; McDougall and Barker, 2011; Lemmon et al, 2007; Connolly, 2009; Balog et al, 2003). In this section, we first review the differentiation states and ocean existence, followed by more detailed descriptions of interior structures of the most prominent ocean worlds among the outer solar system satellites.

1.1 Differentiation and Ocean Existence

Mass and radius allow calculation of the mean density and an assessment of whether a satellite is rich in rock/iron or in ice. The gravity data, in particu-

lar the J_2 and C_{22} components (e.g., Anderson et al, 1996, 1998b,a, 2001), can be used to derive the moment of inertia factor (MoI) if the satellite can be assumed to be in hydrostatic equilibrium. Only for Titan have both gravitational coefficients been determined and this ratio of nearly 10/3 is compatible with hydrostatic equilibrium at 2σ (Durante et al, 2019). Together with mass and radius, the MoI allows construction of simple, albeit non-unique, interior structure models that indicate whether or not a satellite has differentiated. The MoI factor of a homogeneous density sphere is 0.4, and a smaller value indicates an increase of density with depth, hence possible differentiation. Ice on the surface together with a low enough value of the moment of inertia factor allows speculation about a water/ice layer on top of a rock layer. Table 1 collects data on the mass, radii, and MoI of major icy satellites of the solar system.

Table 1 Mass, radii, and moment of inertia of major icy satellites of the solar system. Data: Hussmann et al (2015) for Europa, Ganymede, and Callisto; Iess et al (2014), Roatsch et al (2009), and Jacobson et al (2006) for Enceladus; and Durante et al (2019) for Titan.

	Europa	Ganymede	Callisto	Enceladus	Titan
Mass (10^{22} kg)	4.8	14.8	10.8	0.01	13.5
Radius (km)	1565	2631	2410	252	2575
Mean density (kg)	2989	1942	1835	1609	1881
MoI	0.346	0.312	0.355	0.335	0.341

The bulk densities of Ganymede, Callisto, Enceladus, and Titan suggest that their interiors contain 40 to 60% of ice/water, while Europa is a predominantly rocky body with a bulk ice/water mass fraction of only 6-9% (Hussmann et al, 2015). The level of differentiation of the interiors, however, likely differs between the satellites. Europa and Ganymede are thought to be fully differentiated into a central metallic core, a silicate mantle, and outer water ice-liquid shell (e.g., Anderson et al, 1996, 1998b). Enceladus is differentiated with a water ice-liquid outer shell and central rocky core that may be porous given the satellite’s low mass and mean density, which prohibits a substantial metallic contribution (e.g., Iess et al, 2014; Roberts, 2015; Āadek et al, 2016; Beuthe et al, 2016). In contrast, Callisto and Titan are only partially differentiated with an H_2O layer overlying a core of ice mixed with rocks and metal up to significant depth, maybe up to the center (e.g., Anderson et al, 1998b; Sohl et al, 2003; Iess et al, 2010; Castillo-Rogez and Lunine, 2010; Tobie et al, 2005; Gao and Stevenson, 2013; Baland et al, 2014). A model of slow and incomplete differentiation of Callisto has been discussed by Nagel et al (2004), while Barr and Canup (2010) suggest partial differentiation of Titan due to undifferentiated accretion and core formation due to impacts that allowed only some of Titan’s rock to form a core.

It is widely agreed that these icy satellites have an outer ice I layer that is, in most cases, underlain by an ocean. The strongest observational evidence for the icy Galilean satellites was the detection of magnetic induction signals counteracting the time-variable magnetic field of Jupiter in the satellites’ rest frames (Khurana et al, 1998; Neubauer, 1998; Zimmer et al, 2000; Kivelson et al, 2002). These signals are best explained by the presence of a electrically conducting fluid (i.e. a salty ocean) beneath the surface of the satellite. Strong geologic evidence for a global ocean also exists for Europa (e.g., Pappalardo et al, 1999), and observations

of oscillations in auroral ovals by the *Hubble Space Telescope* have confirmed that Ganymede has a global subsurface ocean (Saur et al, 2015). In contrast, the case for Callisto is less clear since induction within the satellite’s ionosphere may also explain the observed magnetic fields (Hartkorn and Saur, 2017).

For satellites of Saturn, the same approach is not feasible since Saturn’s magnetic field is not inclined with respect to the rotation axis, in contrast to Jupiter (dipole tilt of 9 degrees), and therefore the satellites do not sense a systematic time-periodic field in their rest frame. Instead, the existence of subsurface oceans and characterisation of their properties have relied on a variety of other methods. *Cassini* measurements of water vapour (e.g., Porco et al, 2006; Dougherty et al, 2006) and salty grains from geysers on Enceladus (e.g., Postberg et al, 2009) indicated the existence of water reservoirs beneath the surface, and the global character of the distribution of water as a subsurface ocean was demonstrated by gravity data (McKinnon, 2015) and libration measurements (Thomas et al, 2016). Evidence for a subsurface ocean on Titan is based on the tidal Love number estimation from time-varying gravity field (Iess et al, 2012), detection of an electric perturbation by Huygens during its descent through Titan’s atmosphere that was interpreted as a Schumann resonance (Béghin et al, 2012), and precise measurements of the spin pole orientation (e.g., Baland et al, 2014).

The bottom of the water layer may border to a rocky layer as in the case of Europa and Enceladus or to a high pressure ice layer as is likely the case for Ganymede, Callisto, and Titan. It is also possible that the underlying ice layer is mixed with rock as may be the case for Callisto and Titan if these are incompletely differentiated. High-pressure ice layers are not feasible on Europa and Enceladus due to their small size. More detailed discussions on the the high-pressure ice layer can be found in Journaux et al (2019) (this issue).

The feasibility of subsurface oceans from energy balances of the satellites has been concluded by, for example, Spohn and Schubert (2003), Hussmann et al (2006), and others (compare section 3 below). Maintaining an ocean until the present day requires energy sources and/or the depression of the ice melting point due to the inclusion of other components such as salts or ammonia. Possible energy sources are internal heating coming from radioactive decay in the rocky part of the satellite (e.g., Spohn and Schubert, 2003); dissipation of tidal energy in the rocky interior (e.g., Choblet et al, 2017a), the ocean (e.g., Tyler, 2009; Wilson and Kerswell, 2018), and/or the outer ice shell (e.g., Hussmann et al, 2006); and ohmic dissipation in the ocean (Gissinger and Petitdemange, 2019). Tidal heating is likely important for Europa and Enceladus, but less so for Titan, Ganymede, and Callisto. This is because of Europa’s proximity to Jupiter and the Laplace resonance and because Enceladus likely has a porous core that maintains strong tidal friction; both satellites also have the smallest pressure gradient shifting the water ice triple point to greater depth. Ohmic dissipation is expected to be relatively weak, but can be enhanced locally (Gissinger and Petitdemange, 2019, see Section 4.3). Another crucial factor in sustaining an ocean is heat transport in the ice I layer. Spohn and Schubert (2003) (see also references cited in section 3 below) investigated various scenarios assuming a purely conductive and a convective ice I layer. They find that for pure water ice, convection might lead to complete freezing of the oceans, although the results depend on uncertain parameter values for the viscosity of the ice I layer and the scaling of convective vigor. A present-day global ocean for Enceladus has proven hard to explain based on thermal evolution mod-

143 els, which predict heat production rates given by present orbital conditions below
 144 the expected global heat flow (Roberts and Nimmo, 2008; Tobie et al, 2008). The
 145 discrepancies between estimates of surface heat flow and tidal heating rates could
 146 potentially be explained if Enceladus formed relatively recently, if tidal heating
 147 and cooling were highly time variable rate (episodic or periodic), or if the effective
 148 rate of dissipation within Saturn is larger than the conventional value (Nimmo
 149 et al, 2018).

150 1.2 Internal Structure

151 With this general picture in mind, we will now review more detailed structures for
 152 each of the satellites. Interior models that represent possible internal structures of
 153 Jovian and Saturnian icy ocean worlds are shown in Figure 1.

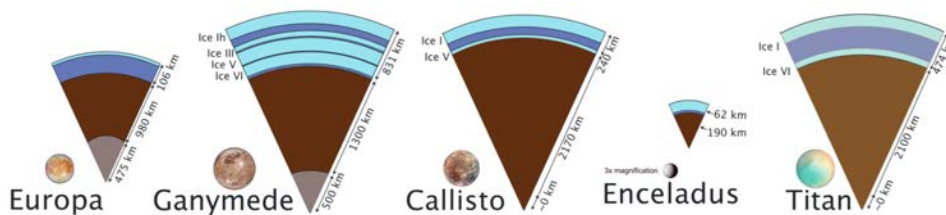


Fig. 1 Spherically symmetric internal structure models that are consistent with geophysical constraints and use state-of-the-art equations of state and thermodynamic properties from Vance et al (2018a). *Cassini* radio science and imaging measurements provide further details for the structures of Enceladus (Čadež et al, 2019; Hemingway and Mittal, 2019) and Titan (Corlies et al, 2017; Durante et al, 2019), revealing that their ice shells are not uniform in thickness, likely owing to thermal or compositional heterogeneities.

154 1.2.1 Europa

155 Gravity data in combination with the mass and radius constraints permit con-
 156 struction of density profiles with radius. These profiles typically take the form of
 157 three-layer interior models an outer ice-liquid water layer, a rocky mantle, and a
 158 central metallic core for Europa. Hydrostatic equilibrium is also assumed because
 159 independent measurements of C_{22} and J_2 are lacking (e.g., Schubert et al, 2004).
 160 For a three-layer model, the core and mantle radii can be determined if the density
 161 of each layer is assumed, leading to uncertainties in their values. In addition, solid
 162 ice and liquid water layers cannot be distinguished due to the small density con-
 163 trast between them. These models suggest that Europa has a outer H_2O layer that
 164 ranges from 80 km to 170 km (Anderson et al, 1998b; Sohl et al, 2002). Geologic
 165 and geodynamic arguments predict ice shell thicknesses that range from ~ 3 km
 166 to >30 km based on mechanical, thermal, cratering, and other methods (Billings
 167 and Kattenhorn, 2005, see their Table 1 for a summary). The core radius depends
 168 on its assumed composition as well as the water layer thickness, ranging from 700
 169 km for a Fe-FeS eutectic core composition and 100 km thick water layer to 200 km
 170 for a pure Fe core composition and 170 km thick water layer (Sohl et al, 2002).

171 Mantle densities are consistent with an olivine-dominated mineralogy, becoming
172 increasingly forsterite rich with decreasing water layer thickness (Sohl et al, 2002).

173 Magnetic field measurements add additional constraints on the interior struc-
174 ture and composition because their observational characterisation can, in principle,
175 constrain the electrical conductivity, depth beneath the surface, and thickness of
176 the ocean (e.g., Zimmer et al, 2000; Khurana et al, 2002; Seufert et al, 2011).
177 Since the *Galileo* mission only observed induction from the main signal caused by
178 Jupiter’s synodic period seen in the satellite’s rest frame, it was not possible to
179 estimate these parameters individually. Schilling et al (2007) found magnetic field
180 data are best explained by electrical conductivity values of $\gtrsim 0.5$ S/m with ocean
181 thicknesses of $\lesssim 100$ km.

182 The composition—dissolved organic and inorganic speciation, salinity, and
183 pH—of Europa’s ocean is poorly constrained. Most models and aqueous leach-
184 ing experiments suggest that magnesium sulfate (MgSO_4) is the dominant salt,
185 in contrast to sodium chloride (NaCl) as in Earth’s ocean (Fanale et al, 2001;
186 Kargel et al, 2000; Zolotov and Shock, 2001; McKinnon and Zolensky, 2003), al-
187 though the concentration varies strongly between models and spans nearly five
188 orders of magnitude across the literature. Recent spectroscopic observations from
189 Earth (Fischer et al, 2015; Trumbo et al, 2019), which trade the higher spatial
190 resolution of *Galileo* near-infrared imaging for better spectral resolution—reveal
191 chlorides associated with active features. Recent interpretations of *Galileo* Near
192 Infrared Mapping Spectrometer (NIMS) data in the light of new laboratory spec-
193 tra find that perchlorates—oxidized Cl ostensibly from the internal ocean—can
194 also match absorption features of surface materials (Hanley et al, 2014). The red-
195 dish tint of Europa’s non-icy materials surface materials has been attributed to
196 radiation-induced flaws in crystalline sodium (e.g., Hand and Carlson, 2015). These
197 many lines of evidence for endogenous chlorine do not rule out a sulfate dominated
198 ocean. As noted by Zolotov and Kargel (2009), a highly oxidized ocean dominated
199 by Mg^{2+} and SO_4^{2-} also has substantial Na^+ and Cl^- . Equilibrium freezing of
200 such an ocean yields a fractionated eutectic composition of mainly NaCl that is
201 nearly identical to the result of applying the same method to seawater (Vance
202 et al, 2019). Further complicating the interpretation of Europa’s ocean compo-
203 sition based on the composition of its surface, the speciation of surface salts is
204 influenced by radiation and by the speed at which freezing (or refreezing) oc-
205 curs (Vu et al, 2016). Thus, it is best to remain agnostic about Europa’s ocean
206 composition until firmer constraints on interior structure, ice thickness, surface
207 composition, and potentially plume composition, can be obtained with NASA’s
208 *Europa Clipper* mission.

209 1.2.2 Ganymede

210 Models for Ganymede’s internal structure, again constrained by mass and gravity
211 data under the assumption of hydrostatic equilibrium, suggest an outer ice-liquid
212 water layer between 600 to 900 km thick, with significant high-pressure ice phases;
213 an intermediate mantle with thicknesses up to 1000 km and density consistent with
214 an olivine-dominated, mostly dehydrated composition; and a central metallic core
215 whose radius may extend from about 500 km to more than 1000 km depending
216 on core composition (Anderson et al, 1996; Deschamps and Sotin, 2001; Kuskov

217 and Kronrod, 2001; Sohl et al, 2002; Vance et al, 2014, 2018a). Multiple pressure-
 218 induced phase transitions are expected within Ganymede’s outer water layer, and
 219 the outermost ice I shell is expected to be less than ~ 150 km thick (Vance et al,
 220 2014, 2018a).

221 Measurements of the induced magnetic field by *Galileo* and auroral oval oscil-
 222 lations observed by the *Hubble Space Telescope* indicate that the ocean electrical
 223 conductivity is at least 0.09 S/m, which corresponds to a minimum salt concen-
 224 tration of 0.9 gram MgSO_4 per kilogram of ocean water, for an ocean between 150
 225 to 250 km depths (Saur et al, 2015). As for Europa, a bias toward a magnesium
 226 sulfate ocean composition for Ganymede is based firstly on models for the aqueous
 227 alteration of CI chondrites (Kargel, 1991)—subsequently shown to be erroneous
 228 (McKinnon and Zolensky, 2003)—that provided a good match to *Galileo* NIMS
 229 spectra that fit well to MgSO_4 . The ocean’s oxidation state, and thus the domi-
 230 nant ionic composition, remain to be confirmed by further measurements, notably
 231 by the European Space Agency’s *JU*piter *IC*y moon *EX*plorer (JUICE) mission
 232 Grasset et al (2013). The intrinsic magnetic field of Ganymede further implies the
 233 formation of an iron-rich core that may itself be layered with a solid inner and
 234 fluid outer core (e.g., Rückriemen et al, 2018).

235 1.2.3 Callisto

236 The interior structure of Callisto is the least constrained of the Galilean satellites.
 237 Here, the interpretations are less clear because the satellite may not be in hydro-
 238 static equilibrium (Gao and Stevenson, 2013) and the induced magnetic field signal
 239 may be due to the ionosphere rather than a subsurface ocean (Hartkorn and Saur,
 240 2017). The MoI-factor assuming hydrostatic equilibrium (Anderson et al, 2001)
 241 prohibits a metallic core, requires low mantle densities that are consistent with a
 242 saturated pyrolite composition, and corresponds to water layers that are $\lesssim 250$
 243 km thick (Vance et al, 2018a). Conversely, a significantly lower MoI estimate that
 244 does not assume hydrostatic equilibrium (Gao and Stevenson, 2013) requires a
 245 central iron core and mantle densities that are consistent with an anhydrous pyro-
 246 lite composition (Vance et al, 2018a). If an ocean is present, Zimmer et al (2000)
 247 found that the magnetic field data are best explained by electrical conductivity
 248 values of $\gtrsim 0.02$ S/m with ocean thicknesses of $\lesssim 300$ km. Because Callisto formed
 249 farthest from Jupiter of the Galilean satellites, any ocean that is present may be
 250 nearly frozen, making it an appealing target for studying the end stage dynamics
 251 of a large ocean world.

252 1.2.4 Enceladus

253 Recent observational data from the *Cassini-Huygens* mission has shed new light
 254 on the interiors of Saturnian satellites. At Enceladus, measurements of the gravity
 255 field, the shape and rotational state, and direct sampling of plume material all pro-
 256 vide constraints on internal structure and composition. Early shape and gravity
 257 measurements, in combination with geyser activity near the south pole, indicated
 258 the presence of a subsurface water reservoir (e.g., Porco et al, 2006; Dougherty
 259 et al, 2006; Thomas et al, 2007; Collins and Goodman, 2007; Iess et al, 2014). A
 260 global ocean was determined decisively by detection of a significant physical libra-
 261 tion (Thomas et al, 2016). The libration amplitude is about four times larger than

262 expected for a solid Enceladus due to the decoupling of the rotational behaviour
263 of the shell with respect to the deeper solid interior and indicates that the ocean is
264 about 20 km beneath the surface on average and that the mean ocean thickness is
265 between 21 and 67 km (Thomas et al, 2016; Van Hoolst et al, 2016). Independent
266 confirmation of these results was obtained by gravity (Iess et al, 2014) and topog-
267 raphy (Nimmo et al, 2011; Tajeddine et al, 2017) data that predict, assuming
268 isostasy at the long wavelengths observed in the gravity field, a core radius of ~ 190
269 km, an ocean thickness of ~ 40 km, and a shell thickness of ~ 20 km, on average
270 (Beuthe et al, 2016; Ćadek et al, 2016). Large variations in ice shell thickness
271 exist, with mean equatorial, north polar, and south polar thicknesses of approx-
272 imately 30 km, 15 km, and 5 km, respectively. (McKinnon, 2015; Thomas et al,
273 2016; Beuthe et al, 2016; Ćadek et al, 2016, 2019; Hemingway and Mittal, 2019).
274 A thinner ice shell at the south pole of Enceladus favors the exchange between the
275 rocky interior, where hydrothermal processes are likely present (Hsu et al, 2015),
276 and the surface (see Section 2.4).

277 Out of all the water oceans that inevitably exist in the universe, Enceladus'
278 is the one that we know about second best (Glein et al, 2018; Postberg et al,
279 2018a). Measurements of the composition of grains and gases erupted out of Ence-
280 ladus in the form of a plume show that the satellite's ocean contains four classes
281 of materials. The first class is salts (Postberg et al, 2009, 2011) that are domi-
282 nated by sodium chloride (NaCl), and sodium bicarbonate (NaHCO₃) or carbon-
283 ate (Na₂CO₃). Potassium salts also appear to be present, but are $\sim 10^2$ times less
284 abundant than their sodium counterparts (Postberg et al, 2009). Second, *in situ*
285 observations of dust in the inner Saturn system (Hsu et al, 2015) indicate that
286 some plume grains from Enceladus contain embedded nanometer-sized particles
287 of nearly pure silica (SiO₂). Third, the major plume volatiles are H₂O, H₂, NH₃,
288 CO₂, and CH₄ (Waite et al, 2017). The presence of minor and trace species, includ-
289 ing volatile organic compounds (VOCs), is also implied by the mass spectrometry
290 data from *Cassini*, although identifying and quantifying individual minor and trace
291 species is challenging because the insufficiently resolved mass spectra allow multi-
292 ple degenerate solutions to the composition (Magee and Waite, 2017). Two effects
293 that may lead to more uncertainty in the volatile composition are chemical reac-
294 tions induced by grain impacts onto instrument surfaces (Waite et al, 2009), and
295 adsorption of VOCs onto ice grains in the plume (Bouquet et al, 2019). These ef-
296 fects can be partially mitigated by focusing on the most weakly adsorbing volatiles
297 (e.g., hydrocarbons) during the slowest flybys. The fourth class of materials in the
298 plume are macromolecular organic compounds (Postberg et al, 2018b). The data
299 from *Cassini* suggest that these materials have high molecular masses (>200 u),
300 and are carbon-rich (low H/C ratios) owing to an abundance of unsaturated car-
301 bon atoms in unfused benzene rings. These features result in hydrophobic phase
302 separation from water. The data also suggest the presence of further chemical
303 complexity in the form of oxygen- and nitrogen-bearing functional groups in the
304 observed organic matter.

305 1.2.5 Titan

306 Multiple lines of evidence from the *Cassini-Huygens* mission also constrain Titan's
307 interior. The observation of a Schumann-like resonance by the *Huygens* probe is

308 not compatible with an entirely solid satellite and can be explained by a conduct-
309 ing ocean 55 to 80 km below the surface, which acts as a reflecting boundary for
310 Extremely Low Frequency (ELF) waves (Béghin et al, 2012). Gravity and topog-
311 raphy also indicate an ocean that is, on average, about 100 km beneath the surface
312 (Nimmo and Bills, 2010). Measurements of the time-varying gravity field of Titan
313 determining the tidal Love number ($k_2 = 0.616 \pm 0.067$) (Durante et al, 2019) also
314 imply an ice shell thickness between 50-100 km (e.g., Mitri et al, 2014), which is
315 consistent with thermal modeling results (Tobie et al, 2006; Mitri et al, 2010).

316 The gravitational constraints provide not only unambiguous evidence of a sub-
317 surface ocean close to the surface, but also indicate that Titan’s subsurface ocean
318 is likely much denser than pure water. The ocean appears to have a high bulk den-
319 sity, exceeding 1100 kg/m^3 , based on the large value of the measured tidal Love
320 number (Iess et al, 2012; Baland et al, 2014; Lefevre et al, 2014; Mitri et al, 2014;
321 Vance et al, 2018a; Durante et al, 2019). Magnesium and ammonium sulfates have
322 been proposed on the basis of chemical and physical models (Fortes et al, 2007)
323 and in the context of experiments investigating the chemistry of these compounds
324 (Hogenboom, 1995; Hogenboom et al, 1997; Vance and Goodman, 2013). The high
325 density constraint could be met with 10 wt% MgSO_4 (Vance et al, 2018a). A re-
326 ducing ocean dominated instead by chlorides can obtain similar large densities,
327 but equation of state data in the relevant pressure range are not yet available to
328 demonstrate this. A saline ocean is not expressly required by the current uncer-
329 tainty in the Love number, though, which permits densities as low as 1100 kg/m^3 ,
330 consistent with the pure water or even 3 wt% ammonia (NH_3) cases as described
331 by Vance et al (2018a). The high Love number could alternatively be explained
332 by a significant viscous behavior of the interior below the ocean (Durante et al,
333 2019) or by a resonantly excited internal gravity mode, which would require the
334 ocean to be stably stratified (Luan, 2019).

335 A dense, salty ocean and thin ice require a high heat flux, exceeding 800 GW
336 (10 mW/m^2) (Vance et al, 2018a). Such a high heat flux would be consistent
337 with recent geological activation of Titan, perhaps concurrent with the formation
338 of Saturn’s rings (Ćuk et al, 2016). A dense ocean and thin ice only worsen the
339 problem of accounting for the low density (2600 kg/m^3) of Titan’s rocky interior.
340 It thus seems likely that Titan is weakly differentiated, highly porous, or both.
341 Alternatively, a differentiated Titan with a small metallic core ($R < 400 \text{ km}$)
342 would be permitted by the gravitational constraints if the low density layer under
343 the ocean can be explained (Vance et al, 2018a). The presence of some dissolved
344 electrolytes in Titan’s ocean solutes is consistent with the model used for the ELF
345 waves, and with a potential low temperature at the top ocean compatible with a
346 likely rigid ice shell (Vance et al, 2018a). An intriguing possibility is that the low
347 density of Titan’s interior can be explained by the presence of organic materials.
348 Geochemical modeling can reproduce the ratios of $^{36}\text{Ar}/\text{N}_2$ and $^{15}\text{N}/^{14}\text{N}$ as mea-
349 sured by *Huygens* in Titan’s atmosphere, if the building blocks of Titan contained
350 abundant organic materials that were subsequently heated and outgassed from
351 the deep interior (Miller et al, 2019). This idea is challenging to model because it
352 requires thermodynamic descriptions of organic-rich mineral assemblages that are
353 rare or non-existent in Earth’s geology.

2 Surface Exchange Processes

Potential surface expressions of ice-ocean material exchange occur vary widely in morphology and age across the satellites. The mechanisms of exchange can be broadly categorized into those caused by impact processes, tectonic processes, cryovolcanic/outgassing processes, and plume processes; a unique morphology characterized by *in situ* surface disruption suggestive of lithospheric thinning, termed chaos, is observed on Europa and will be discussed separately. Europa exhibits global resurfacing with a surface age between 30 and 90 million years (Bierhaus et al, 2009), while Enceladus' surface exhibits both ancient terrains and ongoing geologic activity (e.g., Patterson et al, 2018). Titan's surface is also geologically active, with its thick atmosphere and hydrocarbons seas playing a significant role that will not be discussed in detail here (see Jaumann et al (2009) for a review). Conversely, Ganymede and Callisto show older and more limited signs of surface exchange (Schenk et al, 2004).

2.1 Impact Processes

Multi-ringed structures observed on the icy worlds Callisto, Ganymede, and Europa range in size from the ~ 2000 km diameter Valhalla basin ring system on Callisto (Kinnon and Melosh, 1980; Moore et al, 2004; Schenk et al, 2004) to the ~ 300 km diameter Tyre multi-ringed structure on Europa (Schenk et al, 2004; Schenk and Turtle, 2009). They all share morphological characteristics that suggest impact into a relatively thin brittle lithosphere underlain by a ductile or liquid subsurface (Kinnon and Melosh, 1980; Melosh, 1989). The scale of multi-ring basins on Callisto and Ganymede suggests the possibility of direct exchange between the ice shell and ocean of the satellites. However, the depth to the ice-ocean interface at the time of formation for these basins is not known, leaving open the possibility that their ice shells were not breached during the formation of these features. An abrupt transition from complex crater morphologies to multi-ring morphologies observed on Europa indicates a similarly abrupt transition from ice to water may occur at depths of 10 to a few 10s of km (Schenk et al, 2004) and suggests impacts that formed multi-ring structures could have sampled the satellite's subsurface ocean. However, a lack of radial faulting associated with the formation of multi-ring structures on Europa argues that they may not have breached its ice shell (Turtle, 1998; Kadel et al, 2000). While clear evidence for direct ice-ocean material exchange is not present in association with this process, the potential for convection within the ice shells of Callisto, Ganymede, and Europa (Shoemaker et al, 1982; Schubert et al, 2004; Barr and Showman, 2009) indicates that indirect ice-ocean exchange could still occur.

Palimpsests are impact features that appear to be unique to Callisto and Ganymede. They are generally circular to slightly elliptical albedo features that leave a barely discernable topographic imprint and are characterized by faint concentric lineations and, often, a central smooth region (Schenk et al, 2004; Patterson et al, 2010). Their diameters are measured in 100s of km and, similar to the older multi-ring basins, their formation has been attributed to impact into a relatively thin brittle lithosphere (Shoemaker et al, 1982). As with multi-ring basins, it is possible that direct ice-ocean exchange could have occurred when these features

399 formed, but not clearly so. However, as with the multi-ring basins and structures,
400 indirect exchange of material is also a possibility.

401 Several numerical studies using hydrocodes have been performed to investigate
402 under which conditions melt may be generated upon impact and impact cratering
403 may excavate oceanic water to the surface (e.g., Artemieva and Lunine, 2003;
404 Kraus et al, 2011; Senft and Stewart, 2011). For thin ice shells (≤ 10 km), a pro-
405 jectile of a few kilometers in diameter is sufficient to break the entire shell and
406 expose water to the surface (e.g., Turtle and Pierazzo, 2001; Lunine et al, 2010).
407 For thicker ice shells, exposure of oceanic water is still possible if the projectile
408 size is about half the ice shell thickness (e.g., Artemieva and Lunine, 2005; Lunine
409 et al, 2010). Large impacts such as the one that formed the Menrva crater on
410 Titan, for example, should have brought large volume of water to the surface and
411 temporarily changed the climate of Titan by potentially rising the surface tem-
412 perature by 80 K (Zahnle et al, 2014). Monteux et al (2016) also showed that an
413 impactor of 25 km in radius at moderate velocity (~ 2 km s $^{-1}$) was able to totally
414 disrupt the ice shell and excavate huge volume of oceanic water to the surface.
415 Even if such large impact events remain rare during the moon history, they have
416 the potential to induce resurfacing from regional to global scale, for sufficiently
417 large impacts.

418 2.2 Tectonic Processes

419 Ridges on Europa come in a variety of morphological forms, are observed on length
420 scales of up to 1000s of kilometers, and can range from linear to cycloidal to
421 anastomosing in planform (Prockter and Patterson, 2009). Double ridges are by
422 far the most common ridge type on Europa and are observed over most of the
423 satellite's visible surface history (Figueredo and Greeley, 2000, 2004). Numerous
424 models have been suggested for the formation of ridges on Europa, all of which
425 appeal to the exploitation of a pre-existing fracture in Europa's ice shell. The
426 most widely accepted model of ridge formation suggests that cyclical strike-slip
427 motion on a pre-existing fracture will dissipate heat and cause the warmer, now
428 more buoyant, ice flanking the fracture to uplift and form a double ridge (Nimmo
429 and Gaidos, 2002; Han and Showman, 2008; Kalousova et al, 2016). This process
430 could also create melt that would migrate down the fracture and, provided the
431 fracture penetrates the brittle lithosphere, could provide a direct or indirect path
432 of bringing surface material to Europa's subsurface ocean. The path taken would
433 depend on the thickness and rheology of the shell. Other models of ridge formation
434 suggest they could be pathways for fissure eruptions (Kadel et al, 1998), dike
435 intrusions (Turtle et al, 1998), linear diapirism (Head et al, 1999), or melt squeezed
436 to the surface via cyclical tidal (Greenberg et al, 1998). More recently, subsurface
437 sills feeding cryoclastic eruptions has been proposed (Dombard et al, 2013; Craft
438 et al, 2016). In contrast with the shear heating model, these models imply that
439 ocean material would be brought to the surface or near surface.

440 Some double ridges and ridge complexes (another morphological feature class)
441 on Europa are flanked by deposits that are relatively low albedo and extend for
442 up to 10 km on either side of the feature they are associated with (Lucchitta
443 and Soderblom, 1982; Belton et al, 1996). The dark material is likely a relatively
444 thin surficial deposit that drapes over the preexisting terrain (Geissler et al, 1998;

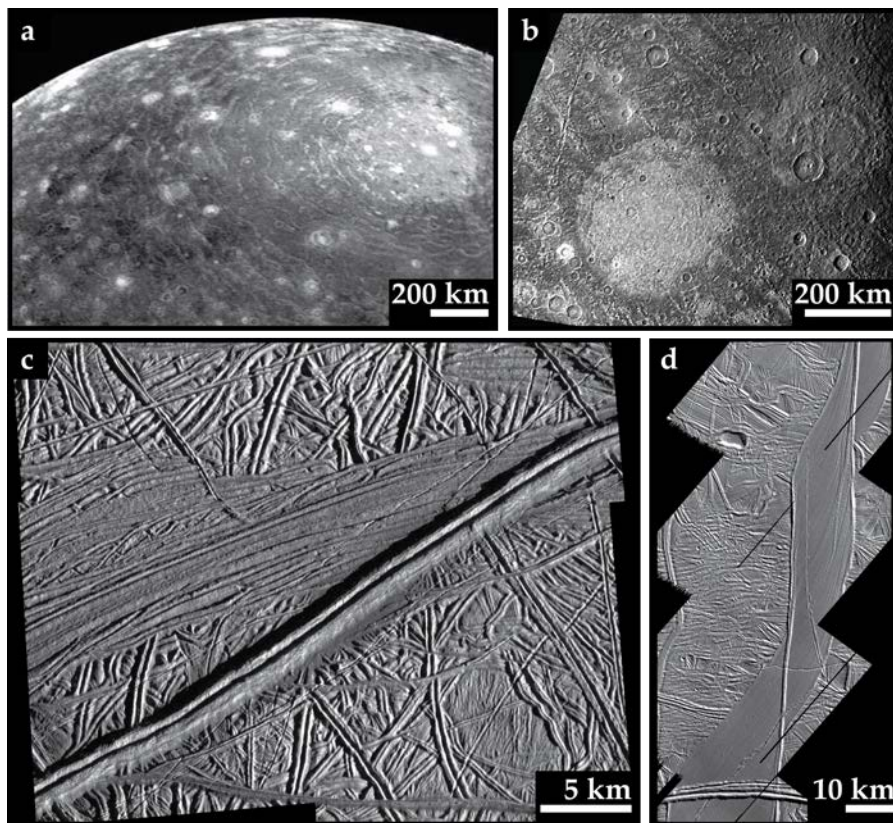


Fig. 2 Impact and tectonic features of icy ocean worlds. (a) The prominent impact basin Valhalla on Callisto. The circular region is 300 km in diameter and basin rings extend 1500 km from the basin center. (b) The 340 km diameter palimpsest Memphis Facula on Ganymede. (c) An archetype double ridge on Europa, Androgeous Linea. (d) A dilational band, Astypalaea Linea, on Europa.

445 Fagents et al, 2000). These deposits may be continuous along the flanks of a ridge,
 446 or spaced in discrete subcircular regions along the margins of a ridge (Prockter
 447 and Patterson, 2009). Observations by the *Galileo* NIMS instrument suggest that
 448 low albedo deposits associated with tectonic features on Europa are composed
 449 of sulfates (McCord et al, 2002) or $MgCl_2$ (Brown and Hand, 2013; Ligier et al,
 450 2016) that are converted into magnesium sulfates through radiolytic processes.
 451 The proposed compositions of low albedo deposits suggest that they were initially
 452 emplaced by an endogenic process and have subsequently been affected by exposure
 453 to the local radiation environment.

454 Ridges on Enceladus also come in a variety of forms (Patterson et al, 2018)
 455 but the most relevant of them for discussing ice-ocean exchange are the ‘tiger
 456 stripes’ of the South Polar Terrain (SPT). The SPT is a pervasively fractured,
 457 geologically young, and low-lying region bound by a quasi-polygonal circumpolar
 458 system of scarps that are intermittently broken by Y-shaped structures (Porco
 459 et al, 2006; Helfenstein, 2014). Within this terrain are the ridges Damascus Sulcus,
 460 Baghdad Sulcus, Cairo Sulcus, and Alexandria Sulcus, collectively referred to as

461 tiger stripes. These features are associated with anomalously high heat flows and
462 are geologically active, as evidenced by eruptive jets (see Section 2.4) of water
463 and other constituents (e.g., Porco et al, 2006; Hansen et al, 2008) that are likely
464 sourced directly from Enceladus' subsurface ocean (Spencer et al, 2018).

465 Bands on Europa are another class of tectonic feature whose formation could
466 facilitate ice-ocean material exchange. This feature can be subdivided into three
467 morphological classes: dilational bands, bright bands, subsumption bands. Dila-
468 tional bands, also referred to as pull-apart bands, are the more commonly observed
469 feature type (Figueredo and Greeley, 2000, 2004). These bands have margins that
470 can be easily reconstructed (Schenk and McKinnon, 1989; Pappalardo and Sulli-
471 van, 1996; Sullivan et al, 1998), indicating that their interiors consist of subsurface
472 material that has been emplaced at the surface of Europa (e.g., Howell and Pap-
473 palardo, 2018). Dilational band formation represents a significant process by which
474 Europa's crust has been resurfaced (Schenk and McKinnon, 1989; Pappalardo and
475 Sullivan, 1996; Prockter et al, 2002). Two endmember models have been proposed
476 for the formation of pull-apart bands. One is the tidal pumping model proposed
477 by (Tufts et al, 2000). That model suggests that bands are part of a continuum
478 process that begins with the formation of a fracture, progresses to a ridge, and
479 ultimately ends in the formation of a dilational band. This mechanism proposes
480 direct exchange of ocean material with the surface of Europa. The second model,
481 by Prockter et al (2002), proposes that band formation is distinct from that of
482 ridges and involves solid-state material rising to fill the separating margins of a
483 preexisting fracture in a manner analogous to terrestrial mid-ocean ridges. This
484 mechanism would imply indirect exchange of ocean material with the surface. Ana-
485 log wax experiments have indicated that oblique opening and shearing commonly
486 associated with the formation of dilational bands is best explained with the latter
487 model of formation (Manga and Sinton, 2004). Bright bands are linear features
488 that disrupt preexisting terrain and have internal textures reminiscent of dila-
489 tional bands. However, unlike that feature type, bright bands are far less common
490 and have margins that do not appear as if they can be reconstructed. Forma-
491 tion mechanisms relying on dilational, contractional, and/or lateral deformation
492 have all been proposed to explain the unique characteristics of this type of band
493 (Prockter and Patterson, 2009). Depending on the formation mechanism used (or
494 combination thereof), ice-ocean exchange is possible, but without additional data
495 to test the proposed formation models, the potential importance of this feature
496 type for material exchange is not as clear as it is with dilational bands. Recent
497 work has introduced a new band feature class: subsumption bands (Kattenhorn
498 and Prockter, 2014). This feature type has been observed within Falga Regio on
499 Europa and is associated with the loss of surface material. It is not clear if material
500 subducted in this manner would reach the ice-ocean interface, though.

501

502 2.3 Cryovolcanic and Outgassing Processes

503 The potential for cryovolcanic activity on Ganymede has changed significantly
504 between analyses conducted using *Voyager* versus *Galileo* data. Based on *Voy-*
505 *ager* images, dark material on Ganymede was interpreted to have been modified
506 by cryovolcanic activity (Murchie and Head, 1989; Croft et al, 1994). This in-

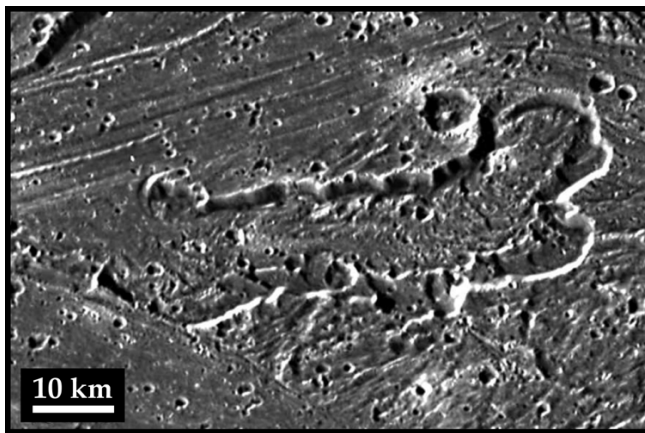


Fig. 3 Oblique view of a depression found within Sippar Sulcus, Ganymede, and acquired during the G8 encounter at 179 m/pixel, where south is up. This feature has a surface texture that may be indicative of flow toward its open end, consistent with it being a source region for icy volcanic material. From (Patterson et al, 2010).

507 terpretation was supported by an apparent absence of small craters, embayment
 508 relationships observed in association with large craters, and smooth areas asso-
 509 ciated with tectonic and impact features (Casacchia and Strom, 1984; Murchie
 510 et al, 1990; Schenk and Moore, 1995). Groove lanes that pervasively disrupt dark
 511 material on Ganymede were interpreted to represent regions of resurfacing by cryo-
 512 volcanic flows, which were subsequently tectonized in some areas to form grooves
 513 (Golombek and Allison, 1981; Golombek, 1982; Shoemaker et al, 1982; Allison
 514 and Clifford, 1987). However, higher-resolution *Galileo* image data imaging of
 515 Ganymede revealed no unequivocal observation of lobate materials with an iden-
 516 tifiable source vent or any other identifiable morphology related to cryovolcanism
 517 associated with dark material (Prockter et al, 2000). Candidate cryovolcanic units
 518 identified from *Voyager* data at lower resolution on the basis of embayment and
 519 texture instead appeared to be the result of fluidized impact ejecta (Pappalardo
 520 et al, 2004) and dark smooth materials in topographic lows appeared to have
 521 accumulated by downslope movement of loose material, instead of by some cryo-
 522 volcanic mechanism (Prockter et al, 1998). Higher-resolution *Galileo* image data of
 523 groove lanes on Ganymede have also lacked clear morphological evidence for flow
 524 fronts, source vents, embayment relationships, or any other evidence suggestive
 525 of cryovolcanic emplacement. However, indirect evidence for volcanic resurfacing
 526 has been identified in the form of small isolated caldera-like features (Lucchita,
 527 1980; Schenk and Moore, 1995; Spaun et al, 2001) and smooth, topographically
 528 low bright lanes (Schenk et al, 2001).

529 For Titan, the only ocean world with a dense atmosphere, evidence of out-
 530 gassing comes from the presence of ^{40}Ar in Titan's atmosphere (Niemann et al,
 531 2005; Waite et al, 2005; Atreya et al, 2006) because ^{40}Ar is produced by the decay
 532 of ^{40}K that is initially contained in the silicate fraction. The amount of ^{40}Ar in
 533 Titan's atmosphere was measured by the Gas Chromatograph Mass Spectrome-
 534 ter (GCMS) onboard the *Huygens* probe in 2005 (Niemann et al, 2005) and by
 535 the *Cassini* Ion and Neutral Mass Spectrometer (INMS) (Waite et al, 2005). The
 536 value was revised to $3.39 (\pm 0.12) \times 10^{-5}$ mole fraction by Niemann et al (2010).
 537 Depending on the elementary composition of the silicate fraction, the outgassing
 538 corresponds to 5 to 20% of the total amount of ^{40}Ar produced by the decay of
 539 ^{40}K . Another clue for the existence of exchange process comes from the presence

540 of methane in Titan’s atmosphere because methane is destroyed by photolysis.
541 Current models (Wilson and Atreya, 2000; Bézard et al, 2014) suggest that the
542 present amount of methane would disappear in less than 30 Myr, which is a short
543 time relative to geological timescales. Therefore, methane has to be resupplied into
544 the atmosphere, and endogenic (e.g., cryovolcanic) processes have been proposed
545 as a possible process (Tobie et al, 2006). Potential cryovolcanic features on Titan
546 are relatively young, not widely distributed across the surface, and include flow
547 fields near Hotei Arcus, Tui Regio, and Ganesa Macula (see Jaumann et al (2009)
548 for a review).

549 2.4 Plume Processes

550 As described in Section 2.2, there are tectonic processes on Enceladus and Eu-
551 ropa that can provide potential (in the case of Europa) or actual (in the case of
552 Enceladus) conduits for ice-ocean exchange relating to plume activity.

553 For Enceladus, approximately 100 supersonic jets of gas and ice grains have
554 been observed to erupt from the four SPT tiger stripes to form a large plume
555 towering above the south pole (Spahn et al, 2006; Porco et al, 2006; Hansen et al,
556 2008; Porco et al, 2014; Spitale et al, 2015). Observed plume intensities vary be-
557 tween 100 kg/s and 1000 kg/s of water vapour, with an average of about 200 kg/s,
558 escaping into the plume (Hansen et al, 2011, 2017). The vapour redeposits onto the
559 vent’s ice walls or condenses to tiny ice grains (e.g., Ingersoll and Pankine, 2010;
560 Schmidt et al, 2008; Yeoh et al, 2015), and a substantial part of the ice grains
561 appear to be frozen ocean spray entrained in the flow that might directly sample
562 the composition of the ocean (Postberg et al, 2009, 2011). Estimates for the gas to
563 ice ratio in the plume vary greatly, although recent estimates suggest an average
564 value of ~ 10 (Kempf et al, 2018; Postberg et al, 2018a). While the ejection speeds
565 for plume vapour are generally above Enceladus’ escape speed (Goldstein et al,
566 2018), only a fraction of the ice grains escape the moon’s gravity to form Saturn’s
567 E ring (Kempf et al, 2018) and a greater part falls back to form surface deposits
568 (Scipioni et al, 2017; Southworth et al, 2019).

569 The jets and plume are temporally and spatially variable. Jets appear to turn
570 on and off on typical time scales of years, indicating occasional opening / sealing of
571 certain ice vents (Nimmo et al, 2014), and systematic variations observed across the
572 fissures suggest trends in the composition of the plume material and/or variations
573 in the plumbing connecting these reservoirs to the surface (Hedman et al, 2018).
574 In contrast, plume activity is coupled most prominently to the moon’s orbital
575 period, with brightness variations on the order of years as well (Hedman et al,
576 2013; Nimmo et al, 2014; Ingersoll and Ewald, 2017); variations in the integrated
577 emitted gas flux over time seem to be milder (Hansen et al, 2017; Teolis et al,
578 2017).

579 The detection of silica nano particles (Hsu et al, 2015), salts, and large or-
580 ganic molecules in the erupted ice grains (Postberg et al, 2009, 2011, 2018b) in
581 combination with MH_4 and H_2 measured in the plume (Waite et al, 2017) sug-
582 gests that material originating from the moon’s rocky core enters the plume. This
583 indicates that the tiger stripe fractures penetrate the entire thickness of the ice
584 shell, tapping into the global ocean underneath (Porco et al, 2006; Kite and Rubin,
585 2016; Spencer et al, 2018). From buoyancy arguments, water should fill large parts

586 of these fractures and the level of neutral buoyancy should be situated at $\sim 90\%$
 587 of the distance from the ocean to the moon's surface, above which the fractures
 588 would be vapor-filled. With an apparent ice shell thickness of not more than 5 km
 589 (e.g., Čadek et al, 2019), it seems plausible that liquid water could be situated at
 590 only a few hundred meters depth within the fractures with some variability (~ 10 s
 591 of meters) due to flushing from tidal flexing of the crust (Kite and Rubin, 2016).
 592 *Cassini* measurements constrain the outlet diameters to be < 10 m (Goguen et al,
 593 2013), and models suggest that the average width of cracks narrows to less than
 594 a few 10s of centimeters above the water surface (Schmidt et al, 2008; Postberg
 595 et al, 2011; Nakajima and Ingersoll, 2016) and is on the order of 1 m for the water
 596 filled portion (Kite and Rubin, 2016; Spencer et al, 2018).

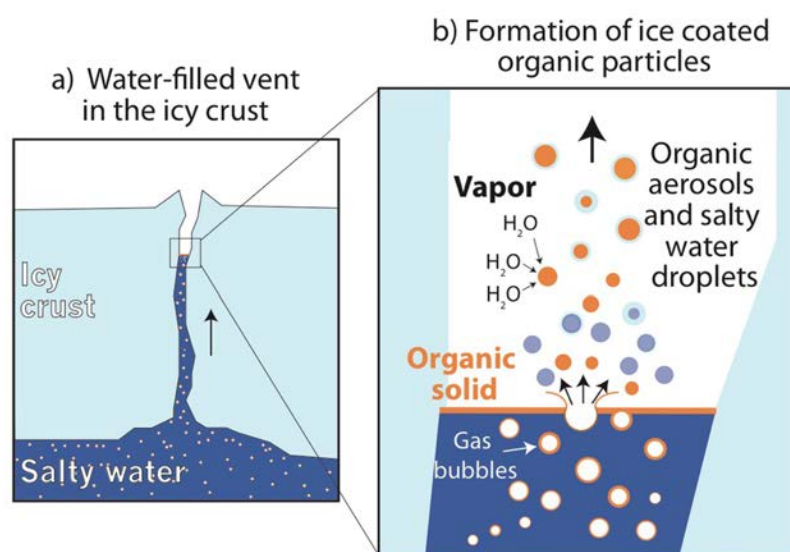


Fig. 4 Schematic of the formation of ice grains from heterogeneous nucleation (not to scale). (a) Ascending gas bubbles in the ocean efficiently transport organic material into water-filled cracks in the south polar ice crust. (b) Organics ultimately concentrate in a thin organic layer on top of the water table inside the icy vents. When gas bubbles burst, they form aerosols made of insoluble organic material that later serve as efficient condensation cores for the production of an icy crust from water vapor, thereby forming organic-rich particles. Another effect of the bubble bursting is that larger, pure saltwater droplets form, which freeze and are later detected as salt-rich ice particles in the plume and the E ring. The figure implies the parallel formation of both organic and saltwater spray, but their formation could actually be separated in space (e.g., at different tiger stripe cracks) or time (e.g., dependent on the varying tidal stresses). From Postberg et al (2018b).

597 The mechanical and thermodynamic driver of the plume is evaporation of ocean
 598 water from the water surface inside back-pressured ice vents (Spencer et al, 2018).
 599 There, temperatures and pressures are close to the triple point of water, which
 600 allows water to evaporate efficiently. Together with volatile gases emerging from
 601 depth or exsolving from the ocean water, vapor is quickly accelerated to supersonic
 602 speeds (Hansen et al, 2008) by the pressure gradient to nearby open space (Gold-
 603 stein et al, 2018). During this ascend, the gases cool substantially and, depending

604 on their composition, will partially condense onto walls and into ice grains (Waite
605 et al, 2017; Bouquet et al, 2019; Khawaja et al, 2019). Almost pure water ice grains
606 and most of the likewise salt poor, but organic-bearing grains are thought to form
607 in this way from supersaturated vapor inside (Schmidt et al, 2008; Postberg et al,
608 2009) and at the outlets (Yeoh et al, 2015) of ice vents. The majority of ice grains
609 in the plume are in a crystalline state (Dhingra et al, 2017), indicative of formation
610 temperatures above 135 K.

611 The apparent heterogeneity of ice grain compositions strongly argues for differ-
612 ent grain formation mechanisms (Fig. 4). Salty ice grains are thought to be frozen
613 ocean spray generated when bubbles burst at the water surface inside the vertical
614 cracks (Postberg et al, 2009, 2011). These bubbles might be formed from either
615 mildly boiling water close to its triple point or upwelling volatile gases (e.g., CO₂,
616 CH₄, or H₂). Consequently, these grains seem to be samples of oceanic near-surface
617 waters (Postberg et al, 2009). A similar mechanism has been proposed to form ice
618 grains containing complex organic substances in high concentrations. In analogy
619 to similar processes on Earth’s oceans (e.g., Wilson et al, 2015), this solid organic
620 material might have accumulated as part of an organic film near the oceanic sur-
621 face. Upon bubble bursting, these organics become aerosolized and then serve as
622 condensation cores to form a water ice crust that is entrained in the vapor flow
623 rising through Enceladus’ ice vents (Postberg et al, 2018b).

624 For Europa, the first tentative telescopic detection of a plume occurred dur-
625 ing a *Hubble Space Telescope* observation in December 2012. Localized ultraviolet
626 line emission of hydrogen and oxygen were attributed as dissociative products of
627 H₂O vapor (Roth et al, 2014). Using off-limb observations while absorbing back-
628 ground light during Europa transits in front of Jupiter, Sparks et al (2016, 2017)
629 twice found indications for a plume at identical positions by. However, both au-
630 thors also report non-detections on several occasions, indicating either sporadic
631 or at least highly variable activity. In a reanalysis of *Galileo* magnetometer data
632 recorded below 400 km altitude during the space crafts closest Europa flyby, Jia
633 et al (2018) reported anomalies consistent with plume activity close to the position
634 of the Sparks et al (2016, 2017) observations. Each of the individual observations
635 does not provide unequivocal proof of a plume. However, the sum of all obser-
636 vations with three different techniques, argue strongly for some level of at least
637 intermittent venting activity. The origin of these putative plumes, however, re-
638 mains an open question. Although a similar interpretation has been invoked for
639 Europa as Enceladus (Southworth et al, 2015), the absence of correlation with true
640 anomaly (Sparks et al, 2017) and the much larger gravity on Europa challenge this
641 interpretation.

642 2.5 Chaos Terrain

643 A terrain unique to Europa, and covering approximately a quarter of its surface,
644 is termed Chaos. Chaotic terrain is formed by disruption of the preexisting sur-
645 face into isolated plates, coupled with the development of lumpy matrix material
646 between the plates. Models for the formation of chaotic terrain that have been pro-
647 posed in the literature fall into 1 of 5 categories – melt-through, diapirism, brine
648 mobilization, sill injection, or impact – and are reviewed in Collins and Nimmo
649 (2009). The melt-through model for chaos formation was born from the visible sim-

650 ilarity of plates in chaotic terrain to terrestrial pack ice (Carr et al, 1998; Greeley
651 et al, 1998). In this model, a heat source at the base of the icy shell facilitates melt-
652 ing of the overlying ice, exposing the ocean below, and leading to the formation
653 of plates equivalent to icebergs that float in a matrix of refrozen ocean material
654 (Greenberg et al, 1999; Thomson and Delaney, 2001). The diapirism model for
655 chaos formation proposes that the morphology of chaotic terrain and pits, spots,
656 and domes (collectively termed lenticulae) represents the surface expression of ris-
657 ing diapirs (Pappalardo et al, 1998; Rathbun et al, 1998; Figueredo et al, 2002;
658 Mével and Mercier, 2007). Such diapirs would develop due to either thermal or
659 compositional buoyancy within the ice shell (Barr and Showman, 2009). In an-
660 other model of chaos formation, Head and Pappalardo (1999) and Collins et al
661 (2000) suggest that the formation of matrix material arises from partial melting of
662 non-water-ice, low-melting-point materials and the mobilization of resulting briny
663 liquids within the ice shell. Another way to deliver liquid into the icy shell of
664 Europa is to inject it directly from the ocean. In this formation model, sills of
665 melt form within Europa’s icy shell from pressurized water injected from frac-
666 tures that penetrate its base (Crawford and Stevenson, 1988; Collins et al, 2000;
667 Manga and Wang, 2007). Here, ice-water interactions and freeze out of the liquid
668 sill describe the unique morphological and topographic characteristics of chaos on
669 Europa (Schmidt et al, 2011). Finally, morphological similarities between chaotic
670 terrain on Europa and terrestrial explosion craters (Billings and Kattenhorn, 2003)
671 have led to the suggestion of an impact origin for the formation of chaos (Cox et al,
672 2008; Cox and Bauer, 2015). In this model, floating plates of the original ice sur-
673 face are preserved in a slushy matrix, filling an irregular hole in the ice left by the
674 explosion crater.

675 **3 Ice Shell Dynamics and Exchange Processes**

676 Exchange processes between the deep ocean and the surface can provide key in-
677 formation about the chemistry and organic content of the ocean, including the
678 chemical processes at work at the rocky core/ocean interface, as has been demon-
679 strated for Enceladus. Similarly important is assessing the downward transfer from
680 the surface to the ocean since surface material may provide compounds, such as
681 oxidants, required to maintain the chemical disequilibrium between the ocean and
682 possible hydrothermal fluids in the rocky core, a process that seems required for life
683 (Hand et al, 2007). Exchange between the ocean and the surface involves transport
684 through the icy shell.

685 **3.1 Thermal State of the Ice Shell**

686 The thickness of the outer ice shell is the principal characteristic that influences po-
687 tential exchanges between the interior and the surface (Chyba and Phillips, 2002).
688 It is controlled by thermal equilibrium between the shell and the subsurface ocean,
689 which depends on how the energy from internal heating (radiogenic and/or tidal)
690 is transported through the ice shell – both conduction and subsolidus convection
691 are suitable heat transport mechanisms (e.g., Spohn and Schubert, 2003; Mitri
692 and Showman, 2005; Tobie et al, 2006). The temperature profile is quite different

693 between a conductive (colder) and a convective (warmer) shell. Moreover, various
 694 studies have shown that the convective processes can be separated into different
 695 regimes (e.g., Moresi and Solomatov, 1995). In the stagnant lid regime, a thick
 696 conductive lid is present on top of the convective layer which effectively slows
 697 down the heat transfer and possibly limits the exchange between the ocean and
 698 the surface, while in the mobile lid regime, which represents plate tectonics, the
 699 lid is thinner and more heat and/or material can be transferred (Fig. 5a).

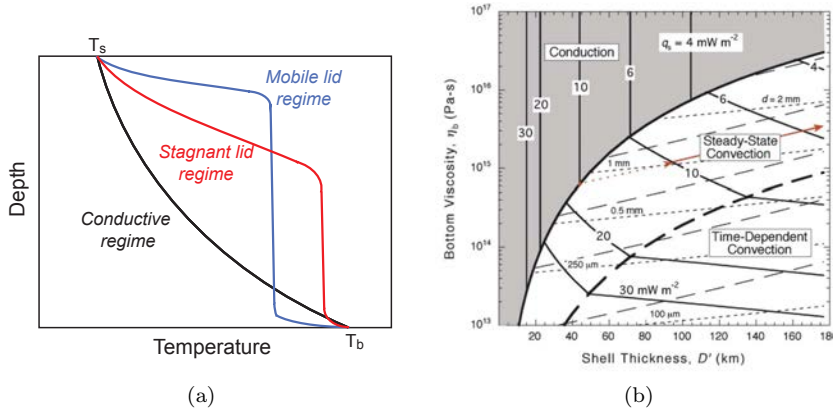


Fig. 5 (a) Conductive (black) and convective thermal profiles in the stagnant (red) or mobile (blue) regime in the ice shell and (b) occurrence of convection (stationary or time-dependent) as a function of ice shell thickness and bottom viscosity in the stagnant lid regime (McKinnon, 2006).

700 Subsolidus convection is an efficient way to transport material between the deep
 701 interior and the surface. On Earth, this process is coupled with plate tectonics that
 702 leads to the major tectonic features such as mid-ocean ridges, subduction zones,
 703 and transform faults. It also produces most of the volcanism. Terrestrial convection
 704 is also characterized by the presence of hot plumes that form at a hot thermal
 705 boundary layer or are triggered by the presence of partial melt (Ogawa, 2014).
 706 Although it is controlled by the same physical processes, thermal convection in icy
 707 shells differs significantly from terrestrial mantle convection for several reasons.
 708 First, silicate mantles are heated from within by radiogenic decay while icy mantles
 709 are mostly heated from below and, in some cases, from within by tidal heating.
 710 Second, internal melting creates a negative buoyancy due to the high density of
 711 liquid water relative to ice, while in silicate mantles, it favors the rise of hot
 712 thermal upwellings. Third, in the case of an icy crust above an internal ocean,
 713 the bottom interface is not fixed as for silicate mantles but evolves depending on
 714 crystallization/melting processes.

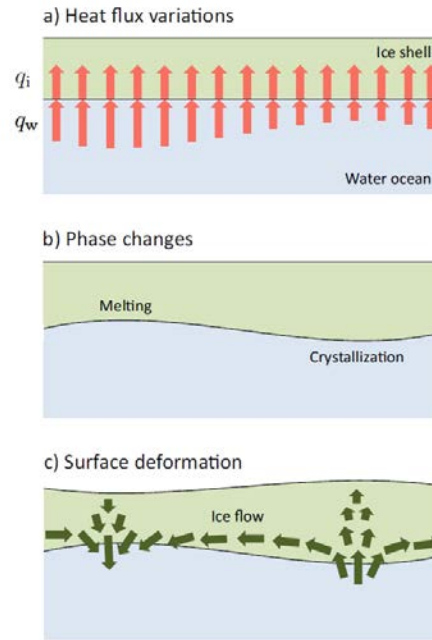
715 Thermal evolution models have provided some constraints on the ice shell thick-
 716 ness of ocean worlds in our solar system although more information is needed to ob-
 717 tain accurate present-day estimates. According to models, the outer ice shells can
 718 undergo large thickness variations during their evolution (Hussmann et al, 2002;
 719 Sotin et al, 2009; Mitri et al, 2010) and potentially produce multiple transitions
 720 between conductive and convective states (Mitri and Showman, 2005). The cou-

721 pling between the thermal and orbital evolution of Europa and Ganymede together
722 with Io in the Laplace resonance could have produced multiple heat pulse events,
723 potentially producing tectonic activity such as grooved terrains on Ganymede and
724 internal melting in the crusts (Showman and Malhotra, 1997; Bland et al, 2009).
725 For most satellites, except probably Enceladus (Choblet et al, 2017a), tidal heat-
726 ing mainly occurs in the ice shell, where the visco-elastic timescale can be of the
727 order of the orbital period of the tidal forcing (Tobie et al, 2003; Sotin et al,
728 2009; Beuthe, 2013). Tidal heating in thin ice shells is larger at the poles than at
729 equatorial regions by a factor of about four (Tobie et al, 2003; Beuthe, 2013) and
730 also strongly depends on regional shell structure (e.g., the presence of faults) and
731 thickness variations, since the tidal stress is approximately inversely proportional
732 to the local shell thickness (Souček et al, 2016; Běhounková et al, 2017; Beuthe,
733 2018; Souček et al, 2019).

734 Although the question of thickness and thermal state of the ice shells is not
735 satisfactorily resolved, various models of solid state convection have been devel-
736 oped. Consolmagno and Lewis (1978) initiated these studies, and more realistic
737 models have subsequently been developed (e.g., Deschamps and Sotin, 2000; To-
738 bie et al, 2003; Barr and Pappalardo, 2005; Mitri and Showman, 2005; Barr and
739 McKinnon, 2007; Han and Showman, 2010; Běhounková et al, 2015; Weller et al,
740 2019) once it was discovered that taking into account the variability of viscosity
741 was key to a good description of heat transfer by convection (Davaille and Jaupart,
742 1993; Moresi and Solomatov, 1995). The vigor of convection is measured through
743 the non-dimensional Rayleigh number with the two main parameters being the
744 ice shell thickness and the ice viscosity. Heat transfer by convection can become
745 more efficient than by conduction as the layer thickens because the conductive
746 heat flux is inversely proportional to the thickness of the layer. The critical value
747 of the thickness at which convection starts has been investigated by a number of
748 studies that included different complexities in the viscosity laws applicable to ice
749 (e.g., Deschamps and Sotin, 2000; Mitri and Showman, 2005; Barr and McKinnon,
750 2007) and showed that the likelihood of subsolidus convection in the icy shells of
751 ocean worlds depends strongly on the deformation properties of ice.

752 The creep behavior of ice I has been studied in laboratory (e.g., Durham and
753 Stern, 2001; Goldsby and Kohlstedt, 2001; De La Chapelle et al, 1999) and ob-
754 served on terrestrial glaciers (e.g., Hudleston, 2015). Several mechanisms can occur
755 to accommodate the deformation rate of ice: propagation of dislocations, diffusion,
756 grain boundary sliding, and basal slip, each of them dominating in certain condi-
757 tions. The deformation rate can be expressed as a combination of these different
758 processes (Goldsby and Kohlstedt, 2001) and depends on pressure (P), temper-
759 ature (T), shear stress (τ), and grain size (d), which also depends on the P - T - τ
760 conditions. Although each process has been well characterized, the grain size is
761 poorly constrained for icy moons (Barr and McKinnon, 2007). Yet its knowledge
762 is crucial to determine the icy shells' thermal state since the smaller the grain
763 size, the lower the value of viscosity and thus the thinner the layer at the onset
764 of convection. Barr and McKinnon (2007) suggest that the minimum thickness for
765 convection to initiate in large ocean worlds such as Ganymede would be between 35
766 and 66 km for grain sizes of 3 to 8 mm, respectively. In a study applied to Callisto,
767 McKinnon (2006) proposes maps showing the conditions for convection to exist
768 as a function of the viscosity at the ice/ocean interface and the thickness of the
769 ice layer (Fig. 5b). Although the domain for convection seems large and suggests

Fig. 6 Sketch of the processes occurring at the ice/water interface and their consequences for the ice shell evolution (Kvorka et al, 2018). (a) The normal component of heat flux is not generally continuous at a phase interface. Melting occurs in regions where heat flux from the ocean is larger than heat flux in the ice shell. The opposite situation leads to crystallization. (b) In regions of melting, the ice shell loses mass and its thickness decreases. Crystallization is accompanied by ice mass gain and ice shell thickening. Both processes tend to restore the heat flux balance at the boundary. (c) Undulations of the ice/water interface generate pressure gradients that induce flows in the ice shell. These flows deform the upper boundary of the shell. Depressions develop above the regions of melting, while the surface tends to rise above the regions of crystallizations.



770 that convection would be dominant for thicknesses larger than 20 km, note that
 771 the value of the corresponding grain size is less than 1 mm, which is more than
 772 on order of magnitude smaller than the values predicted in Barr and McKinnon
 773 (2007). Note also that the grain size evolution model predicts an increase in grain
 774 size after convection starts, leading to increased viscosity and less vigorous convec-
 775 tion. Finally, the presence of impurities would affect the grain size of ice as it has
 776 been observed and modeled in cold ice sheets on the Earth (Durand et al, 2006).
 777 Convection processes in the icy crust may have been intermittent on Ganymede,
 778 Titan (Tobie et al, 2005, 2006), Europa (Hussmann et al, 2002), and Enceladus
 779 (Barr, 2008), and a definite answer has to await measurements by future missions.

780 3.2 Global Dynamics of the Ice Shell

781 Long-wavelength topography and gravity can be used to constrain the lateral vari-
 782 ations in shell thickness. Its amplitude and pattern provide insights on the thermal
 783 state and global dynamics of the ice shell as well as on the coupling with the un-
 784 derlying ocean (e.g., Nimmo et al, 2011). In combination with heat production
 785 within the ice shell, strong heat flux anomalies coming from the seafloor and heat
 786 flux patterns due to oceanic circulation can lead to a modulation of the ice/ocean
 787 interface. The 3D structure of the ice shell and its global dynamics thus result from
 788 a balance between the heat transfer through the ice shell, melting/crystallization
 789 processes at the base and within the ice shell, and lateral ice flow (Čadek et al,
 790 2017; Kvorka et al, 2018)(Fig. 6).

791 From the inversion of the topography and gravity data collected by *Cassini*,
 792 maps of ice shell thickness have been inferred by several studies on Titan (Lefevre

793 et al, 2014; Mitri et al, 2014; Kvorka et al, 2018) and Enceladus (Čadek et al,
 794 2016, 2019; Beuthe et al, 2016; Hemingway and Mittal, 2019). On Titan, the long-
 795 wavelength topography is associated with small gravity anomalies indicating a
 796 high degree of compensation (e.g., Durante et al, 2019). The observed topography,
 797 characterized by relatively small amplitudes (about 1 km peak to peak) and an
 798 anomalous equatorial bulge (the poles are about 300 m lower than the equator)
 799 can be explained either by a deflection of the ocean/ice interface (Nimmo and
 800 Bills, 2010; Hemingway et al, 2013; Lefevre et al, 2014) or by density variations
 801 in the upper crust, likely due to heavy hydrocarbon clathrates (Choukroun and
 802 Sotin, 2012). Assuming that surface topography is due to ice/ocean interface de-
 803 flection, the inferred deflection amplitude (± 5 km) indicates a very slow ice flow
 804 at the base of the ice shell. It also implies a conductive and highly viscous ice
 805 shell above a relatively cold ocean ($T < 250$ K) (Lefevre et al, 2014; Kvorka et al,
 806 2018). By modeling the shape evolution of Titan’s ice shell including diffusive heat
 807 transfer through the ice shell, heterogeneous tidal heating in the ice shell, heat flux
 808 anomalies from the ocean and basal ice flow, Kvorka et al (2018) show that the
 809 observed topography is not consistent with tidal heating pattern in the ice shell
 810 and rather indicates heat flux anomalies in the ocean (Fig. 7a). The anomalous
 811 topographic bulge would be consistent with lateral variations of ocean heat flux
 812 on the order of 0.1-1 mW/m², characterized by upwelling of warm water in polar
 813 regions and downwelling of cold water at low latitudes (Kvorka et al, 2018). The
 814 origin of this oceanic flow is unclear. It may be a natural flow for the present-day
 815 ocean configuration or it may result from heat flux forcing from the base of the
 816 ocean (Soderlund, 2019).

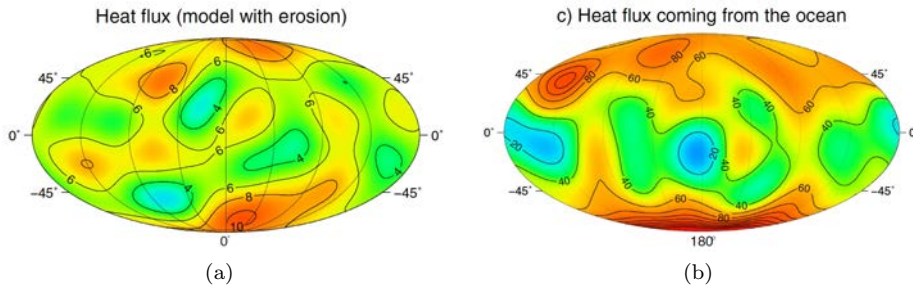


Fig. 7 Heat flux anomalies from the ocean derived from the ice shell thickness variations, assuming a conductive ice shell, on (a) Titan for a model including surface mass redistribution by erosion (Kvorka et al, 2018) and (b) Enceladus (Čadek et al, 2019).

817 Using a similar approach, Čadek et al (2019) estimated the heat flux anomalies
 818 at the bottom of Enceladus’ ice shell in order to explain the observed topography
 819 (Tajeddine et al, 2017) and gravity (Iess et al, 2014). Compared to Titan, the ice
 820 shell thickness variations are much larger, ranging from 5 km at the south pole to
 821 35 km at the equator that are associated with heat flux anomalies about ten times
 822 larger than on Titan (Fig. 7b). By modelling the ice flow driven by variations in
 823 hydrostatic pressure on the ice/water interface, Čadek et al (2019) demonstrated
 824 that Enceladus’ ice shell is in a steady state, with melting located in polar regions
 825 and crystallization occurring in the equatorial region. The observed pattern is

826 consistent with the heat flux pattern predicted by tidally-heated water flow in the
 827 porous core of Enceladus (Choblet et al, 2017a), likely modulated by oceanic flow
 828 (Soderlund, 2019).

829 The global shape data of Europa retrieved from *Galileo* limb profiles indicate
 830 that, if variations exist, they should be relatively small, thus implying an efficient
 831 ice flow at the base of the shell (Nimmo et al, 2007) or redistribution of topogra-
 832 phy through pressure-induced melting and freezing (Soderlund et al, 2014). The
 833 absence of significant ice shell thickness variations does not imply that there is
 834 no significant heat flow anomaly at the base of the ice shell. As shown by Čadek
 835 et al (2017), any ice/ocean deflection relaxes much faster in Europa’s conditions
 836 than in Enceladus’ case because of its larger size. For the same ice shell thickness
 837 and viscosity structure, the relaxation rate is 100 times faster on Europa than
 838 on Enceladus, requiring a large heat flux anomaly to build up significant long-
 839 wavelength topography. Ashkenazy et al (2018) modeled the global meridional ice
 840 flow in Europa’s ice shell due to pressure gradients associated to equator-to-pole
 841 ice thickness variations. They show that the thickness variations barely exceed a
 842 few kilometers and are limited by ice flow and oceanic heat transport.

843 3.3 Melt Transport

844 While the observations of plumes at Enceladus and Europa support the possibility
 845 of liquids present within their ice shells, the physical processes leading to near-
 846 surface melting and liquids accumulation within the crust or their emplacement
 847 on the surface are still subject to discussion. Generation of liquids within the
 848 icy crust requires a heat source, such as tidal heating, and/or the presence of a
 849 secondary phase depressing the melting point. Two geodynamical contexts have
 850 been proposed for generation of melts by enhanced tidal heating inside the ice shell:
 851 either in hot upwelling plumes as a result of thermally-reduced viscosity (e.g., Sotin
 852 et al, 2002; Tobie et al, 2003; Běhouňková et al, 2010) or along the faults due to
 853 tidally-activated strike-slip motions (e.g., Gaidos and Nimmo, 2000; Nimmo and
 854 Gaidos, 2002). Alternative sources of liquids would be a direct connection from
 855 the subsurface ocean through water-filled cracks (Fig. 8a, Section 2.4) and impact
 856 cratering (Fig. 8b, Section 2.1).

857 Contrary to the analogous setting on the Earth (silicate solids being denser
 858 than their melts), the negative buoyancy of water with respect to ice I is often seen
 859 as an obstacle to maintaining englacial water (Tobie et al, 2003; Kalousová et al,
 860 2016). However, several possibilities have been proposed to overcome the negative
 861 buoyancy of water (e.g., Fagents, 2003; Hammond et al, 2018): (i) volatiles such
 862 as CO₂, CO, or SO₂ may be exsolved in water thus significantly increasing the
 863 fluid buoyancy, (ii) non-ice substances may be present in water or ice which will
 864 modify the density contrast between the two phases - either by decreasing the fluid
 865 density (e.g., NH₃, CH₄, N₂) or by increasing the ice density (silicate particles,
 866 clathrates), (iii) compaction and associated low permeability of ice that allows an
 867 accumulation of melts within the shell, and (iv) partial freezing of a discrete liquid
 868 reservoir will lead to its overpressurization, which may further promote cracking
 869 and lead to water ascent.

870 Over long timescales, oceanic materials – either in the form of melt pockets or
 871 solid phase (salts, clathrate) – can be advected from the ice/ocean interface to the

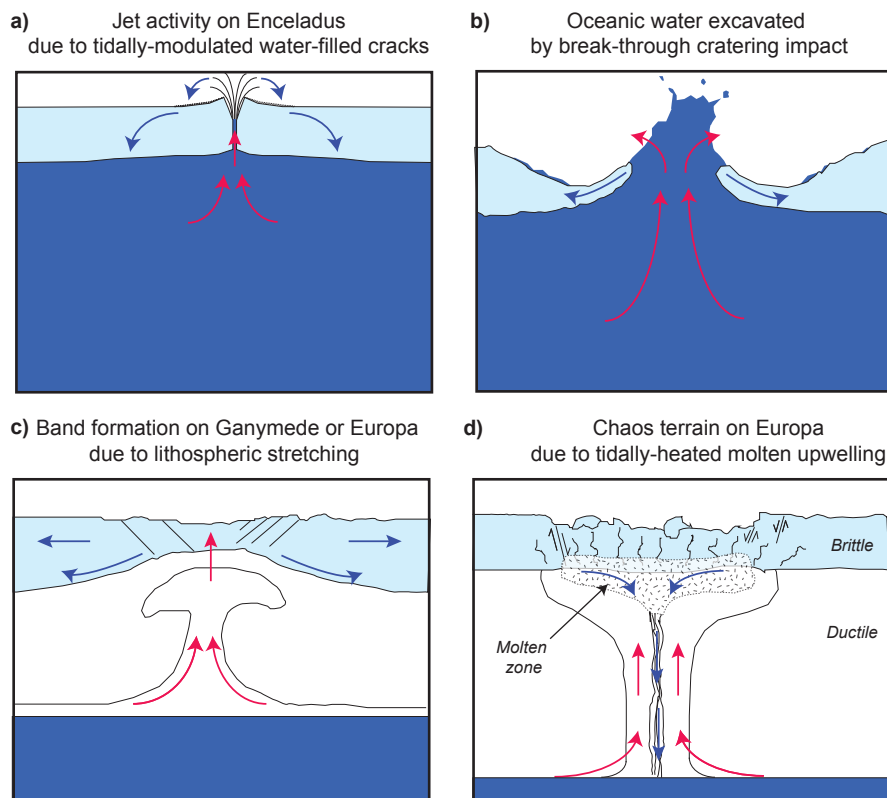


Fig. 8 Four possible mechanisms leading to exchange between the subsurface ocean and the surface, which are mostly controlled by the ice shell thickness and the force acting on it : a) oceanic water injection and associated jet activity requiring relatively thin shells ($\lesssim 5$ km) and strong tidal forces (after Postberg et al, 2018b); b) oceanic water excavation due to break-through cratering impact requiring ice shell thickness and impactor radius to be of comparable size (after Lunine et al, 2010); c) indirect transport of oceanic materials by upwelling ices from the ocean/ice interface to shallow depths, associated with band formation and lithospheric stretching, occurring in 10-20 km thick shells under the action of significant tectonic stress (after Howell and Pappalardo, 2018); d) ice melting at shallow depths by tidally-heated thermal plumes and potential percolation of meltwater to the ocean, occurring for relatively thick ice shells ($\gtrsim 25$ km) subjected to significant tidal forcing (after Sotin et al, 2002).

872 surface by thermal convection. The conductive lid separating the convective part
 873 from the surface (Fig. 5a) acts as a barrier for chemical exchanges. Rupture of the
 874 conductive lid, either by large-scale tectonic stresses and melt-induced collapse
 875 (Fig. 8c-d), is required to allow the exposure of materials brought by convective
 876 upwellings. Rupture of the lid provides also a mean to recycle surface materials
 877 to the subsurface and potentially to advect them to the ocean (Kattenhorn and
 878 Prockter, 2014; Johnson et al, 2017; Klaser et al, 2019). In numerical simulations
 879 of thermal convection in icy shells, melting of water and its transport is often
 880 completely neglected (e.g., Han and Showman, 2005, 2010) or highly simplified

881 (Běhounková et al, 2012). Some authors, however, have included water generation
882 and considered its dynamic effect on the ice flow (Tobie et al, 2003; Kalousová
883 et al, 2016). In these studies, the water content (porosity) is computed but the
884 ice is considered to be effectively impermeable to the interstitial water transport
885 (percolation) and water is thus simply advected by the flowing ice. These authors
886 found that the occurrence of a few percents of water leads to fast (with respect
887 to convection time scales) destabilization of the partially molten region, thus not
888 supporting the long term stability of liquid water bodies. Let us note however,
889 that only pure ice was considered and that the addition of salts may promote
890 the melting process while the addition of volatiles may improve the buoyancy of
891 the liquid. Alternatively, the water transport by interstitial percolation has been
892 modeled by using a two-phase mixture formalism by Kalousová et al (2014) for
893 Europa and Hammond et al (2018) for Triton. The drawback of these studies is that
894 they only consider a one-dimensional geometry and thus neglect water advection
895 by flowing ice (cf. above). The only simulations that took into account both water
896 transport mechanisms, i.e. advection by ice and interstitial percolation, have been
897 performed for the high-pressure ice layers of Ganymede (Kalousová et al, 2018;
898 Kalousová and Sotin, 2018) and Titan (Kalousová and Sotin, 2019). More details
899 can be found in Journaux et al (2019) (this issue). On the basis of the effect of
900 melt on the dynamics and structure of the high-pressure ice layer, one can predict
901 that a temperate layer (i.e. with temperature following the melting curve) would
902 be present at the icy crust/ocean interface but details on its characteristics must
903 await dedicated work.

904 Models have also been constructed to assess the possibility of cryolava ascent
905 on the surface of icy satellites. Manga and Wang (2007) found that overpressure
906 generated by freezing a few kilometers of ice is not sufficient for liquid water to
907 erupt on the surface of Europa, while it may be sufficient on smaller satellites such
908 as Enceladus. However, Quick and Marsh (2016) and Quick et al (2017) found
909 that cryolava may reach Europa's surface at temperatures as high as 250 K and
910 undergo rapid cooling to form cryovolcanic domes. Even if the liquids do not reach
911 the surface, they may be placed in the shallow subsurface where processes such as
912 impact cratering can transport the deposited material on the surface.

913 The presence of near-surface liquid reservoirs was proposed to explain some
914 ocean world surface morphologies in comparison with Earth-like processes. Michaut
915 and Manga (2014) and Manga and Michaut (2017) investigated thermo-mechanical
916 constraints on the emplacement and evolution of liquid water sills and proposed
917 that Europa's pits, domes, and small chaos morphology could result from the evo-
918 lution of these sills located at depths of 1 to 5 kilometers. Moreover, they suggest
919 that the pits should be located above bodies of liquid water, which is in agreement
920 with Schmidt et al (2011) who proposed that Europa's chaos terrains form above
921 liquid water lenses ~ 3 kilometers below the surface. Walker and Schmidt (2015)
922 investigated the effects of a subsurface liquid water body on the flexural response
923 of an ice shell and the resulting topography. Their results reproduce the observed
924 geology of Europa's chaos terrains as well as Enceladus' SPT, suggesting that they
925 both formed by the ice collapse above a liquid water body. Similarly, sills and dikes
926 have also been implicated for the formation of ridges on Europa (Dombard et al,
927 2013; Johnston and Montési, 2014; Craft et al, 2016).

928 4 Ocean Dynamics and Exchange Processes

929 Oceans are an essential component of the ice-ocean exchange process as the inter-
930 mediary layer between the outer ice shell and the underlying mantle/high pressure
931 ice layer. Moreover, because the oceans are nearly inviscid and strong currents
932 are expected, heat and materials are transported relatively quickly across them.
933 Fluid motions within icy satellite oceans are driven by convection due to thermo-
934 compositional density gradients, mechanical forcings (e.g., tides, libration, and
935 orbital precession), and magnetic forcing due to electromagnetic pumping. The
936 resulting flows will promote mixing within the bulk ocean, which will influence
937 the distribution of thermo-compositional gradients, especially along the seafloor
938 and ice-ocean interface, and potentially have important implications for the ice
939 shell and habitability.

940 4.1 Convection

941 4.1.1 Hydrothermal Plumes

942 The earliest efforts to understand ocean dynamics within icy worlds focused on
943 local circulation driven by hydrothermal plumes upwelling from the rocky interior
944 (Thomson and Delaney, 2001; Goodman et al, 2004). The goal of these efforts was
945 to explore connections between localized seafloor heating and geological features
946 on the surface of Europa’s ice crust. Initial work in this area (Thomson and De-
947 laney, 2001) argued that Coriolis forces would constrain the outward spread of
948 the turbulent buoyant plume, creating a narrow “chimney” of warm fluid which
949 could potentially deliver hydrothermal heat to a narrow patch of ice despite the
950 depth of the ocean, and speculated that they could form the 5-20 km diameter
951 lenticulae (Greenberg et al, 1999) or pits, domes, and spots (Pappalardo et al,
952 1998) commonly seen on Europa’s surface. They further noted that these forces
953 would create anticyclonic (counter-clockwise in the northern hemisphere) currents
954 at the ice-water interface, which were consistent with the apparent motion of ice
955 rafts in the large Conamara Chaos (Spaun et al, 1998).

956 Later work based on theoretical scaling laws for point-source plumes (Fernando
957 et al, 1998), supplemented with laboratory tank experiments (Goodman et al,
958 2004) and numerical simulations (Goodman and Lenferink, 2012), demonstrated
959 that while the plumes would have a narrow aspect ratio, they would still be at least
960 20-50 km in diameter, far wider than the common sizes of pits, domes, and spots,
961 suggesting that these features were more likely created by internal ice dynamics.
962 However, the plume size was found to be compatible with the largest chaos regions
963 on Europa. Temperature anomalies were found to be very small (mK) and flow
964 velocities very weak (mm/s), making it unlikely that hydrothermal plumes would
965 have any direct mechanical effect on the overlying ice layer. Mantle heat transport
966 calculations by Lowell and DuBose (2005) also support the lack of melt-through
967 events by hydrothermal plumes.

968 More recent work by Farber and Goodman (2014) included a better treatment
969 of planetary rotation in these plumes and studied their implications for astrobi-
970 ology. In Earth’s oceans, the horizontal component of the planetary rotation axis
971 (perpendicular to gravity) can be ignored, but this approximation is not valid for

972 convection in deep icy world oceans. As a result, hydrothermal plumes should rise
973 parallel to the planetary rotation axis (i.e. diagonally), rather than vertically. This
974 would cause a plume’s projection onto the ice surface to be an ellipse with the long
975 axis oriented toward the pole. Hydrothermal plumes are also astrobiologically sig-
976 nificant, both in providing a route for metabolically significant molecules to move
977 from seafloor to surface, and for delivering potential biosignatures to the ice. Far-
978 ber and Goodman (2014) also showed that tracer particles require thousands of
979 years to move from top to bottom of Europa’s ocean.

980 One important caveat remains, however. Local convective plumes will rise
981 through a background ocean whose properties are determined by the ocean com-
982 position and global circulation. Compositional modification of these results is dis-
983 cussed below, and global convective flows are discussed in Section 4.1.2.

984 The extent to which hydrothermal plumes will buoyantly rise depends on the
985 ocean composition. In a freshwater ocean, buoyancy of the plumes depends on its
986 temperature relative to the surrounding ocean water; since the warmer plumes are
987 less dense, they are expected to reach the ice-ocean interface. However, the thermal
988 expansion coefficient of water is negative between 0 and 3.98 °C (at 1 bar) such that
989 the maximum fluid density is reached below at a temperature above the freezing
990 point. As a result, the ocean would have a stable “stratosphere” beneath the ice-
991 ocean interface that would prohibit further rising of the plume (Melosh et al,
992 2004). Increased pressure and ocean salinity move the temperature of maximum
993 density towards the freezing temperature, reducing this effect. These temperatures
994 coincide at pressures exceeding 27 MPa and salinities exceeding ~ 3 wt% for both
995 seawater and magnesium sulfate compositions (Feistel and Hagen, 1995; Melosh
996 et al, 2004; Vance and Brown, 2005). Further, considering a saline ocean, if the salt
997 content is larger near the seafloor due to interactions with the underlying mantle
998 or precipitates, its entrainment into the plumes would increase their density and
999 cause them to reach the point of neutral density before reaching the ice-ocean
1000 interface (Vance and Brown, 2005). However, Travis et al (2012) argue that this
1001 initial salinity gradient would eventually become homogenized such that thermal
1002 buoyancy would regain dominance.

1003 *4.1.2 Global Circulations*

1004 Heat flow from the seafloor combined with heat loss through the overlying ice
1005 shell is expected to drive thermal convection globally in the oceans, modulated by
1006 compositional buoyancy associated with salinity gradients that may enhance the
1007 vigor of convection (positive gradient) or have a stabilizing effect (negative gradi-
1008 ent). The resulting fluid flows are governed by the Navier-Stokes equations where
1009 the most prominent forces are inertia, Coriolis, pressure gradient, and buoyancy,
1010 in combination with the ocean geometry and boundary conditions (see Taubner
1011 et al (2019), this issue). The hydrothermal plume studies above generally assume
1012 the ocean to be geostrophic, meaning a balance between the Coriolis and pres-
1013 sure gradient forces that effectively organizes the convective flows into quasi-two-
1014 dimensional structures that are aligned with the rotation axis, frequently called
1015 Taylor columns (Fig. 9a). On a global scale, non-linear stresses associated with
1016 these columns will drive an eastward jet at low latitudes with multiple, alternat-
1017 ing jets towards the poles, reminiscent of the jets in Jupiter’s atmosphere (Fig. 9b;

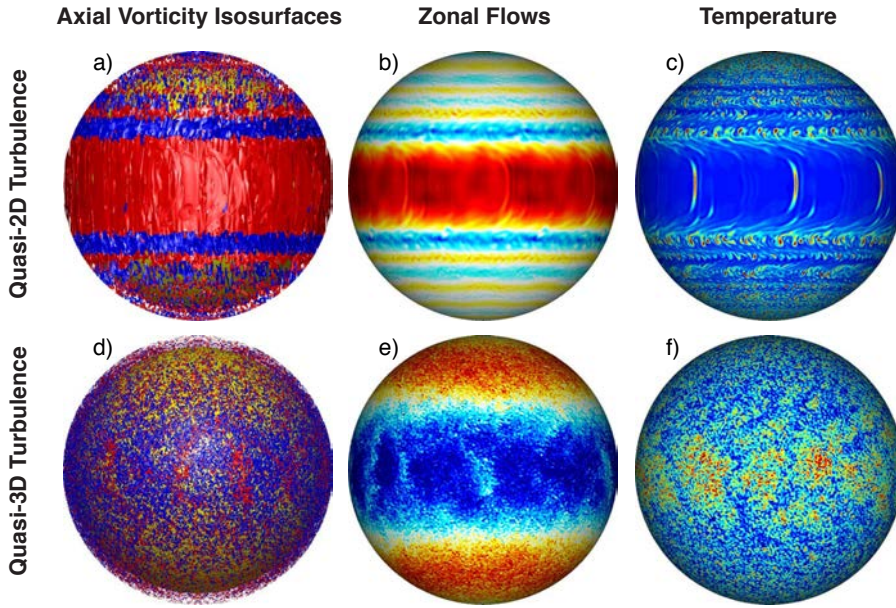


Fig. 9 Flow structures, zonal flows, and temperature fields in convection models at a snapshot in time. Top row: Simulation with quasi-2D turbulence. Bottom row: Simulation with quasi-3D turbulence. Left column: axial vorticity isosurfaces, $\omega_z = (\nabla \times \mathbf{u}) \cdot \hat{\mathbf{Z}}$. Red (blue) indicates cyclonic (anticyclonic) circulations aligned with the vertical $\hat{\mathbf{Z}}$ direction; the yellow sphere represents the seafloor. Middle column: zonal flows along the outer boundary; red (blue) indicates eastward (westward) flow. Right column: superadiabatic temperature fields below the outer boundary; red (blue) indicates warm (cool) fluid. Adapted from Soderlund et al (2014).

1018 e.g., Heimpel et al, 2015; Soderlund et al, 2014). The ocean is warmest at high lat-
 1019 itudes due to the efficiency of vertical convection columns there (Fig. 9c; Aurnou
 1020 et al, 2008).

1021 As convection becomes more vigorous, however, the Taylor columns break down
 1022 and fluid flows become three-dimensionalized. Here, mixing of absolute angular
 1023 momentum will instead drive a westward equatorial jet and eastward jets at high
 1024 latitudes (Fig. 9d; e.g., Aurnou et al, 2007; Gastine et al, 2013). A large-scale
 1025 meridional overturning circulation also develops with upwelling flow near the equa-
 1026 tor and downwelling flow at higher latitudes in each hemisphere (Fig. 9e; Soderlund
 1027 et al, 2013). This circulation brings warm ocean water preferentially toward the
 1028 ice-ocean interface at low latitudes ((Fig. 9f; Soderlund et al, 2014).

1029 The transition between these convective regimes is an active area of research
 1030 subject to considerable debate (e.g., Gastine et al, 2016; Cheng et al, 2018). Al-
 1031 though the oceans are traditionally assumed to be strongly organized by rotation,
 1032 Soderlund et al (2014) postulated that Europa’s ocean is characterized by quasi-3D
 1033 turbulence by estimating the convective regime following several potential scaling
 1034 laws. These arguments are updated and extended to the oceans of Enceladus, Ti-
 1035 tan, and Ganymede in Soderlund (2019), who predicts thermal convection in the
 1036 oceans of Europa, Ganymede, and Titan to all behave similarly (Fig. 9, bottom
 1037 row). Rotation is predicted to play a more significant role for Enceladus. Consid-
 1038 ering the satellites collectively, peak zonal flow speeds are predicted to reach at

1039 least 10s of cm/s (up to meters per second) and mean (i.e. averaged over both
1040 time and all longitudes) radial flows span the mm/s to cm/s range (Soderlund,
1041 2019).

1042 Complementary to the formation of individual chaos features through hy-
1043 drothermal plumes, global convection models have focused the distribution of
1044 chaos terrains across Europa. Soderlund et al (2014) hypothesized that the low
1045 latitude enhancement of ocean heating promotes the formation of these terrains
1046 through increased melting of the ice shell and subsequent accretion of relatively
1047 pure marine ice. Considering the other satellites, heat flux anomalies from the
1048 ocean derived recently for Enceladus (Čadek et al, 2019) and Titan (Kvorka et al,
1049 2018) provide useful constraints for models of these oceans (see Fig. 7). The mod-
1050 els of Soderlund (2019) are consistent with these heat flow patterns only if salinity
1051 effects are taken into account. However, the distribution of heating from the un-
1052 derlying mantle/high pressure ice layer may be spatially heterogeneous (Travis
1053 et al, 2012; Choblet et al, 2017a,b; Kalousová et al, 2018) and these effects have
1054 not yet been taken into account.

1055 Ocean composition and its thermodynamic properties may have a significant
1056 impact on global circulations due to the presence of a stable stratosphere if the
1057 thermal expansion coefficient is negative (Melosh et al, 2004) or if salinity gra-
1058 dients are maintained across the ocean. If salinity increases towards the seafloor,
1059 the thermal and compositional gradients oppose each other and double-diffusive
1060 convection may be expected (Vance and Brown, 2005; Bouffard et al, 2017) with
1061 layering that can evolve into a ‘staircase’ configuration with well-mixed layers
1062 that are characterized by steps in salinity and temperature (e.g., Schmitt, 1994).
1063 Conversely, if salinity increases toward the ice-ocean interface, both thermal and
1064 compositional gradients are unstable, leading to more vigorous convection. Melting
1065 and freezing along the ice-ocean interface will also lead to regions that are locally
1066 enhanced with fresher (i.e. stably stratified) and saltier (i.e. unstably stratified)
1067 water, respectively, that may drive additional circulations (Jansen, 2016; Ashke-
1068 nazy et al, 2018; Zhu et al, 2017). Moreover, if heterogeneous melting/freezing
1069 leads to large-scale topographic variations along the ice-ocean interface, they may
1070 also impact the characteristics of convection and promote mechanically driven
1071 flows (see Section 4.2).

1072 4.2 Mechanical Forcings

1073 The icy satellites are tidally locked, meaning that their rotational periods equal
1074 their orbital periods. This results in the moons having triaxial ellipsoid shapes with
1075 their longer axes pointing towards the planet. If the moons’ orbits were perfectly
1076 circular and their rotational axis aligned with their orbital axis, there would not
1077 be any mechanical forcing. However, their orbits are eccentric and their rotational
1078 axes are tilted with respect to their orbital axes, resulting in time-changing tidal
1079 bulges, librations, and precessions (Fig. 10) that can drive ocean currents.

1080 A common approach for studying ocean tides is to use the Laplace Tidal Equa-
1081 tions (LTE) that control the barotropic ocean response (Hendershott, 1981). The
1082 LTE assume a shallow ocean of constant density, and radial (vertical) ocean cur-
1083 rents are considered to be negligible with respect to horizontal currents such that
1084 the problem becomes 2D. The resulting equations, which allow for surface gravity

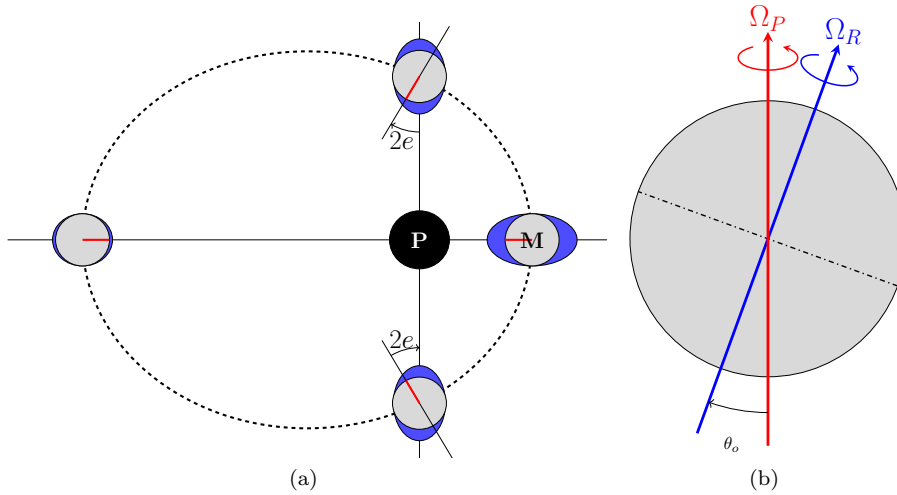


Fig. 10 (a) A moon’s eccentric orbit, where the equilibrium tidal bulge is indicated in blue and the 0° meridian is marked with a red line. Note the change of the tidal bulge amplitude and the longitudinal libration of the subplanet point. (b) A moon’s rotation axes, where the moon rotates around its rotational axis with angular velocity Ω_R and precesses with angular velocity Ω_P , separated by the moon’s obliquity θ_o .

1085 waves and planetary Rossby waves (e.g., Longuet-Higgins, 1968), have been used
 1086 to study the response of an ice-free ocean of constant thickness (e.g., Tyler, 2008,
 1087 2009; Chen et al, 2014; Hay and Matsuyama, 2017). The ocean response highly
 1088 depends on the surface gravity wave speed. For thick oceans, the surface gravity
 1089 wave speed is high, the ocean quickly adjusts to the perturbing tidal potential, and
 1090 its response is mainly given by the equilibrium tide; this is the case of tides raised
 1091 by a satellite’s eccentricity (Fig. 11a). However, a high surface gravity wave speed
 1092 does not hamper the propagation of tangentially non-divergent Rossby waves as
 1093 they do not involve up and down motions (Fig. 11b).

1094 Obliquity of the satellite can excite planetary Rossby waves of sufficient am-
 1095 plitude to maintain a liquid ocean in Europa (Tyler, 2014), but is insufficient to
 1096 prevent Enceladus’ ocean from freezing (e.g., Chen and Nimmo, 2011; Matsuyama,
 1097 2014). For thin oceans, gravity wave resonances can also occur. Nevertheless, char-
 1098 acteristic ocean thicknesses for which these resonances occur are far from those
 1099 inferred from observations; as an example, the thickest ocean for which a resonance
 1100 occurs in Enceladus is around 350 m. If an ocean eventually begins to freeze out,
 1101 it will necessarily go through resonant states where enhanced heat production pre-
 1102 vents further freezing. Recently, Beuthe (2016); Matsuyama et al (2018); Hay and
 1103 Matsuyama (2019) considered the effect of the overlying ice layer and showed that
 1104 obliquity-forced dissipation is enhanced but the eccentricity tide is significantly
 1105 dampened in satellites with high effective rigidity (Enceladus) and enhanced in
 1106 satellites with low effective rigidity (Ganymede, Europa, Titan).

1107 The LTE hold as long as the ratio of the characteristic vertical and horizontal
 1108 length scales is small (Miles, 1974). However, this may not be sufficiently accurate
 1109 for icy satellite oceans. Using the ocean thickness and body’s radius as a measure
 1110 of vertical and horizontal length scales, higher ratios are obtained for Europa and

1111 Enceladus than for Earth (~ 0.06 and ~ 0.15 versus ~ 0.001), suggesting that $3D$
1112 effects are relevant in the icy moons. Without the shallow water approximation,
1113 internal inertial waves can be excited. These waves have properties markedly differ-
1114 ent from shallow water waves. Upon reflection, an internal wave packet can change
1115 its wavelength. Depending on the ocean geometry, this can lead to the focusing of
1116 energy along internal shear layers (e.g., Rieutord et al, 2011; Maas, 2005).

1117 Recently, Rovira-Navarro et al (2019) used the linearized Navier-Stokes equa-
1118 tions to study the three dimensional response of an unstratified ocean of constant
1119 thickness to tidal forcing. They observed patterns of periodic inertial waves that
1120 take energy from the global tidal forcing and focus it along internal viscous shear
1121 layers that propagate in the ocean (see Fig. 11c). These shear layers fluid flows can
1122 have an amplitude of a few cm/s, but the dissipation due to inertial waves in an
1123 ocean of constant thickness is not sufficient to prevent an European or Enceladan
1124 ocean from freezing. Requier et al (2019) extended this work to study the excita-
1125 tion of inertial waves by Enceladus' libration and concluded that this mechanism
1126 generates more dissipation than tidal forcing.

1127 Rovira-Navarro et al (2019); Requier et al (2019) ignored the advection terms in
1128 the Navier-Stokes equations which otherwise can result in flow instabilities and the
1129 development of turbulence. Flow instabilities in spherical and ellipsoidal containers
1130 have been widely studied (e.g., Malkus, 1994; Kerswell and Malkus, 1998; Rieu-
1131 tord, 2004). In a librating sphere, the viscous boundary layer at the solid-liquid
1132 interface can become unstable and break down to small scale turbulence. Wilson
1133 and Kerswell (2018) estimated the amount of tidal dissipation due to boundary
1134 layer instabilities and suggested that it should be potent enough to explain Ence-
1135 ladus' heat flux. If the container is ellipsoidal, the interaction of inertial waves
1136 with the mean-flow excited by libration, precession, or tides can lead to the well-
1137 known elliptical instability, which also results in a turbulent flow regime similar
1138 to that shown in Fig. 11d (e.g., Kerswell, 2002; Le Bars et al, 2015). Experimen-
1139 tal and numerical work shows that Enceladus and Europa are likely unstable to
1140 libration-driven elliptic instability, while Titan, Ganymede and Callisto are prob-
1141 ably not (Grannan et al, 2014, 2017; Lemasquerier et al, 2017). So far, ocean
1142 currents excited by mechanical forcing and convection (Section 4.1.2) have been
1143 studied separately. The interaction of mechanically driven and convective currents
1144 requires further attention.

1145 The previous discussion assumes the ocean to be unstratified and contained
1146 within ellipsoidal or spherical shells devoid of topographical features. On Earth,
1147 stratification and ocean topography play a crucial role in shaping the ocean's
1148 response to tides by controlling the barotropic ocean response and the conversion
1149 of the barotropic tide to the (internal) baroclinic tide (Munk, 1997; Egbert and
1150 Ray, 2000). In icy ocean worlds, we expect the ocean floor and the basal ice shell
1151 topography to deviate from the idealized shapes explored so far in mechanically
1152 driven flow studies. For instance, Enceladus' ice shell thickness varies from ~ 5 km
1153 at the south pole to ~ 30 km at the equator (Section 1.2). Additionally, under
1154 certain circumstances a subsurface ocean might be stratified (Section 4.1.2). The
1155 study of mechanically-excited flows for complex ocean geometries and stratified
1156 subsurface oceans is an exciting topic for future research.

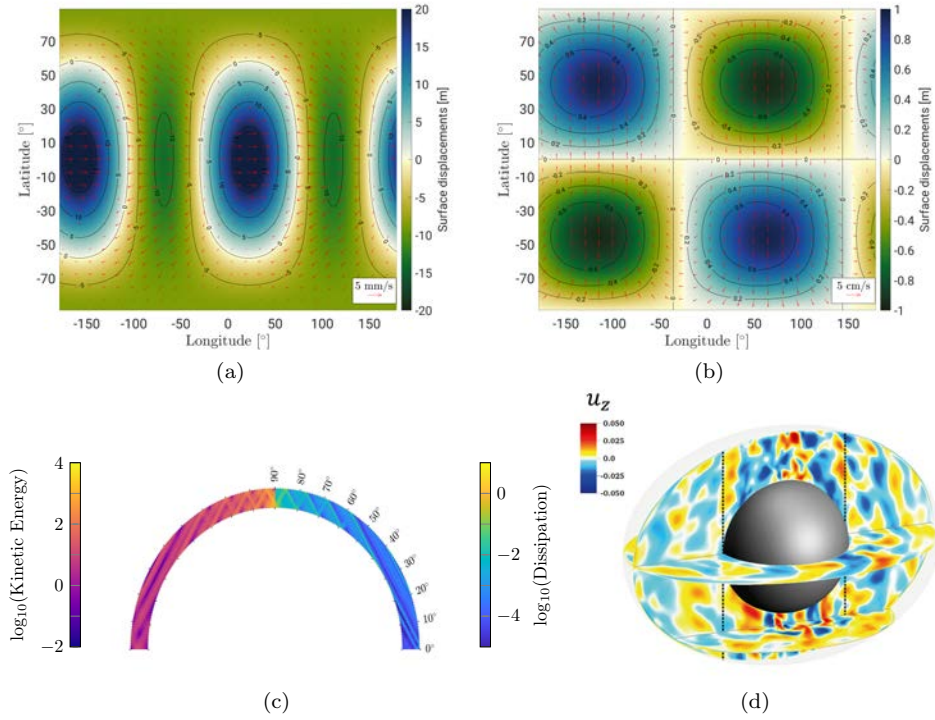


Fig. 11 Characteristic flow fields for mechanical forcings. Flow field and surface displacements excited in Europa by the (a) eccentricity and (b) obliquity tide. The ocean is assumed to be 80 km and covered by a 20 km thick ice shell. Overlaid are contours corresponding to the equilibrium tide. The flow pattern propagates towards the east and west for the eccentricity and obliquity tides, respectively. Credit: M. Rovira-Navarro. (c) Internal waves excited by the eccentricity tide in an ice-free 30 (185) km thick Enceladan (European) ocean; the left and right quadrants respectively show the amplitude of kinetic energy and viscous dissipation. A logarithmic non-dimensional scale is used for both quantities with the maximum kinetic and viscous energy corresponding to 4.7 (0.65) J/m^3 and 0.50 (0.026) $\mu\text{W}/\text{m}^3$ for Enceladus (Europa), respectively (Rovira-Navarro et al, 2019). (d) Turbulent flow regime attained due to the libration-driven elliptical instability, where the vertical component of the non-dimensional velocity (u_z) is shown. For more details, see Lemasquerier et al (2017).

1157 4.3 Magnetic Forcing

1158 Jupiter's magnetic field is offset by $\sim 10^\circ$ with respect to the orbital plane of
 1159 the Galilean satellites. As a result, the satellites experience a time-varying mag-
 1160 netic field that induces electrical currents in the ocean. If the phases of the Jo-
 1161 vian magnetic field and the oceanic induced magnetic field are not aligned, their
 1162 offset could generate a Lorentz force that drives a retrograde jet and weaker up-
 1163 welling/downwelling motions at low latitudes (Gissinger and Petitdemange, 2019).
 1164 Mean flow speeds of the equatorial jet are expected to reach a few cm/s in Europa's
 1165 ocean, reducing to a few mm/s for Ganymede and < 1 mm/s for Callisto where
 1166 the Jovian magnetic field is weaker. This process is not expected to be significant
 1167 for the Saturnian satellites due to the planet's nearly axisymmetric magnetic field.
 1168 Magnetic pumping leads to Ohmic dissipation within the oceans of $\lesssim 10^8$ W (at

1169 Europa), which is several orders of magnitude weaker than both radiogenic and
1170 tidal heating. However, the dissipation may still be significant if it is spatially
1171 concentrated at high latitudes in a thin layer below the ice-ocean interface due to
1172 the skin effect (Gissinger and Petitdemange, 2019). Moreover, because the zonal
1173 jet is characterized only by retrograde flow, it may contribute to the reorientation
1174 of Europa’s ice shell (i.e. non-synchronous rotation) and the associated formation
1175 of lineaments (e.g., Helfenstein and Parmentier, 1985).

1176 5 Terrestrial Analogs

1177 While a great amount of the work on ocean worlds has focused on geologic pro-
1178 cesses analogous to those seen within the solid Earth, studies of ice and ocean
1179 processes on Earth are equally important for understanding physical and biologi-
1180 cal processes on these moons. Here, we will show how analogs for the freezing and
1181 geochemical properties of planetary ice shells can draw from knowledge of sea ice
1182 and marine ice on Earth (e.g., Buffo, 2019) as well as models of Snowball Earth
1183 (e.g., Ashkenazy et al, 2018).

1184 5.1 Terrestrial Ice-Ocean Interfaces

1185 The outer shells of icy satellites likely formed through top-down freeze-out of their
1186 oceans. On Earth, the majority of thick ice (ice sheets/shelves) is meteoric, form-
1187 ing via the compaction of snow. However, sea ice and marine ice on Earth form
1188 directly from the ocean and thus are analogous to what is expected on ocean bear-
1189 ing satellites. While forming under different environmental pressures, the thermal
1190 gradient is the main controller of ice chemistry in both sea ice and marine ice (e.g.,
1191 Buffo et al, 2018; Buffo, 2019). Analogously, it is likely that the ice-ocean inter-
1192 faces of ice shells will be in either a high or low thermal gradient state, depending
1193 on shell thickness and age, where the high thermal gradient regime is similar to
1194 that observed in sea ice and the low thermal gradient by marine ice formed at the
1195 base of ice shelves.

1196 Representative of the high thermal gradient regime, sea ice is easy to observe,
1197 both in the field and with remote sensing, and thus much is known about its struc-
1198 ture and formation. The majority of sea ice is composed of granular and columnar
1199 ice (Dempsey and Langhorne, 2012; Dempsey et al, 2010) driven by turbulent
1200 and quiescent ice-ocean interface conditions, respectively. However, in ice-shelf-
1201 adjacent sea ice in Antarctica, another process occurs that may be relevant to icy
1202 satellites: platelet ice accretion. Through a process known as the ice pump (e.g.,
1203 Lewis and Perkin, 1986), deep ice is melted at high pressures, creating a plume
1204 of fresher water that rises buoyantly along the shelf and out below the sea ice. As
1205 the plume rises, depressurization causes the water to become supercooled (having
1206 temperatures below its *in situ* freezing point) and ice crystals called platelets form
1207 in the water column, rise up to the ice surface, and form a layer of poorly organized
1208 and highly porous ice. Under ice shelves, this layer can grow to immense thick-
1209 nesses (>100 m) and becomes marine ice (Fricker et al, 2001; Craven et al, 2005,
1210 2009; Galton-Fenzi et al, 2012). Under sea ice, this platelet layer can be incorpo-
1211 rated into the growing sea ice or remain unconsolidated. At the base of actively

1212 forming first year sea ice with low current velocities, transient brine drainage into
1213 supercooled water can additionally form brinicles (icicles that form around brine
1214 drainage channels), but these become inactive once the sea ice growth slows or
1215 platelet accretion takes over.

1216 Two lessons can be gained from sea ice and applied to icy satellites. First,
1217 thermal gradients are critical to the structure and composition of the ice. The
1218 thermal gradient within forming columnar sea ice is ~ 10 K/m, for which the
1219 salinity is ~ 4 -5 ppt (freezing from 34 ppt ocean), and it exhibits a critical porosity
1220 of 4-5% beyond which no brine drainage is observed (e.g., Dempsey et al, 2010;
1221 Golden et al, 2007). High thermal gradient conditions are only relevant for very
1222 shallow ice (< 1 km) on icy satellites, or if ocean water is injected rapidly from
1223 depth into the upper regions of their ice shells (Buffo, 2019). In areas where rapid
1224 ice growth does occur, there could be important gradients in rheological properties
1225 that might make these regions more probable to fracture or re-melt (Buffo, 2019).
1226 Second, ice accretion forced by melting of deep draft ice affects the thickness
1227 and properties of distant ice, which is key to consider as an interface process for
1228 other ice-ocean worlds. The thickness of the platelet ice layer is determined not
1229 by ice formation rate through surface cooling, but by the conditions of the ocean
1230 – very thick in large supercooling plumes to non-existent where the ocean is not
1231 supercooled. This layer can be efficiently modeled as an upward sedimented layer
1232 of platelets whose crystal size is controlled by the degree of supercooling and the
1233 layer thickness by the lifetime of the plume (Buffo et al, 2018).

1234 Representative of the low thermal gradient regime, marine ice is found in areas
1235 where supercooled water drives ice accretion onto the base of ice shelves (10s of
1236 meters to kilometers thick meteoric ice). This regime is characterized by very low
1237 thermal gradients, $\ll 1$ K/m. The thickest marine ice observed on Earth is ~ 500 m
1238 thick, comprising about half the thickness of the Amery ice shelf at its midpoint
1239 (Fricker et al, 2001). Borehole observations showed a many tens of meters thick
1240 unconsolidated platelet layer forming at the bottom of the marine ice (Craven
1241 et al, 2005). The marine ice in Amery demonstrates nearly complete brine rejection
1242 with salinities of 0.03-0.56 ppt (Craven et al, 2009), suggesting compaction driven
1243 desalination. Observations of accreting columnar marine ice at the bottom of the
1244 Ross Ice Shelf can be reproduced using sea ice models and lower thermal gradients:
1245 5% critical porosity at 0.08 K/m yields a theoretical salinity of 1.7 ppt, while 1.95
1246 ppt is estimated using constitutive equation based model results (Buffo, 2019),
1247 which agrees well with field observations of 2.32 ppt (Zotikov et al, 1980).

1248 The thermal gradient expected for most ice shells is squarely within the low-
1249 thermal gradient regime (e.g., 0.02 K/m for Europa (McKinnon, 1999)). Once an
1250 ice shell reaches its diffusive limit (< 1 km thick ice), the bulk salinity is unlikely
1251 to appreciably change, meaning that while the ice still contains some salt through
1252 accretion, the properties at the ice ocean interface do not change significantly with
1253 regards to the rate or properties of ice accreted (Buffo et al, 2019), in the absence
1254 of platelet ice accretion. This diffusive limit marks how much ocean material the
1255 ice could deliver to the upper shell through convection or diapirism. These values
1256 are robust if the critical porosity is the physical limit past which brines are trapped
1257 within the ice, which is the observed limit in sea ice (Golden et al, 1998, 2007)
1258 and brine layers in ice shelves (Kovacs and Gow, 1975) and matches observations
1259 of sub-ice shelf columnar ice (Buffo, 2019).



Fig. 12 Platelet ice accreting under the McMurdo Ice shelf, imaged from Icefin (Credit: B. E. Schmidt/Icefin/RISEUP). These ice crystals form in supercooled water below the shelf and accrete onto the ice shelf base. In some places, these compact to become marine ice, which may also be a process relevant to planetary ice-ocean interface processes.

1260 Preliminary work by Buffo (2019) suggests that multiphase, hydraulically con-
1261 nected layers at the base of planetary ice shells are likely a stable, and thus
1262 common, phenomenon. Furthermore, the environmental pressures these layers are
1263 subject to (i.e., gravity, thermal gradient, ocean composition) likely dictate their
1264 thicknesses and structural properties. As the exchange boundary for energy and
1265 mass between the underlying ocean and ice shell, the structure and dynamics of
1266 these regions will substantially impact the thermochemical evolution of planetary
1267 cryospheres. Additionally, in the analogous terrestrial environments (sea ice and
1268 marine ice), the porous nature of the ice-ocean interface provides a gradient rich
1269 substrate that supports a diverse biome.

1270 5.2 Snowball Earth

1271 At least two extreme glaciations occurred on Earth during the late Neoproterozoic
1272 era (750-580 Ma) (Hoffman and Schrag, 2002). The Snowball Earth hypothesis
1273 proposes that the oceans froze over entirely during these episodes, so that the
1274 Earth was an icy ocean world. While debate continues over whether ice cover was
1275 total or partial (Liu and Peltier, 2010), this episode may provide our best terrestrial
1276 proxy for ice covered oceans on a global scale. As with icy worlds in the outer solar
1277 system, a Snowball Earth ocean would be geothermally heated but isolated from
1278 direct solar heating or wind forcing. However, differential solar heating would still
1279 create variations in thickness which would drive global ice flow from pole to equator
1280 (Goodman and Pierrehumbert, 2003). This flow would be balanced by melting and
1281 freezing into the ocean, leading to variations in salinity which would drive ocean
1282 circulation, would close the glaciological cycle through the ocean, creating an ocean

haline-driven circulation (Ashkenazy et al, 2013). All of these dynamics are similar to predictions for the ice-covered oceans of the outer solar system, particularly the coupled interaction of global ice flow and haline-driven ocean circulation (Collins and Goodman, 2007; Ashkenazy et al, 2018; Zhu et al, 2017). Interestingly, the unsolved question “Was the liquid ocean ever exposed at the surface?” is of key importance to both fields.

However, there are three key differences. First and most obviously, Earth’s oceans are relatively shallow and are interrupted by continents. Their depth lessens the importance of planetary rotation (especially its horizontal component; Farber and Goodman, 2014). The continents block global east-west flows of the type explored by Soderlund et al (2014). On the other hand, coastlines allow fluid parcels to change their vorticity, enabling strong north-south currents that would otherwise be limited by angular momentum constraints (Pedlosky, 1987).

A second difference between Snowball Earth and the icy worlds of the outer solar system is that Snowball Earth’s surface temperatures would have been far warmer, leading to a totally different pattern of ice flow dynamics. On Earth, modern-day and Snowball ice sheets flow via uniform strain, gradually spreading and thinning while maintaining constant flow velocity with depth (Weertman, 1957; Goodman and Pierrehumbert, 2003). However, this model does not work in the outer solar system: ice near the surface is too cold to flow at all. Instead, ice moves by vertical shear flow: warm ice near the base of the ice shell flows horizontally beneath a rigid upper shell (e.g., Collins and Goodman, 2007; Ashkenazy et al, 2018).

Third, Snowball Earth would have had much thinner ice than the icy ocean worlds. Models of the Snowball Earth ice thickness typically predict ice 200 - 1000 m thick (Goodman and Pierrehumbert, 2003), though local regions where clear ice allows solar penetration could be as thin as a few meters (Warren et al, 2002). In contrast, ice shell thicknesses in the outer solar system range from ~ 5 km at Enceladus’ SPT to hundreds of kilometers for large satellites (e.g., Vance et al, 2018a). Many of these moons are then likely to be experiencing solid-state convection, while heat flow through Snowball Earth’s ice shell probably occurred by conduction only. The thickness and weak thermal gradients within ice shells of ocean worlds also promotes the horizontal movement of ice via vertical shear rather than uniform strain, as discussed above.

Other important differences also abound, including greater uncertainties about solutes and seafloor heating distribution in the oceans of the outer solar system, but the similarities provide ample ground for collaboration between investigators in both fields.

6 Discussion and Perspectives

6.1 Habitability of Icy Ocean Worlds

The potential habitability of ocean worlds is an exciting prospect that has been commonly discussed for the past three decades, with increasing sophistication (e.g., Reynolds et al, 1983, 1987; Chyba, 2000; Chyba and Phillips, 2001; Hand et al, 2009; Barge and White, 2017; Russell et al, 2017; Schmidt, 2019), mirroring progress in astrobiology towards understanding the requirements to maintain

1328 a habitable planet. During this time, the direction of progress has been towards
1329 systems science, which is important for ocean worlds given that while they share
1330 many similarities with the Earth, conditions may be quite different. In particular,
1331 capturing coupled interactions between the geophysical and geochemical evolu-
1332 tion of planets is needed given that the Earth and its biosphere coevolved (e.g.,
1333 Des Marais et al, 2008; Hays et al, 2015).

1334 Understanding the energy to support life in a given planetary system is the
1335 central organizing principle of the study of habitability. For the ocean worlds, this
1336 is a chance to understand not just what inventory of material they may have, but
1337 which processes continually supply energy for life. Here, the state of the interior,
1338 implied thermal and chemical evolution, and modern exogenic processes all couple
1339 together. Since light transmission even through pure glacial ice ceases at < 10 m
1340 (Christner et al, 2014), photosynthesis is not a viable energetic pathway for the
1341 ocean worlds. Thus, understanding potential pathways for chemosynthesis (Zolotov
1342 and Shock, 2003, 2004; Russell et al, 2017; Barge and White, 2017) is critical for
1343 understanding the habitability of icy ocean worlds.

1344 Europa's global ocean, coupled with a potentially reducing interior (e.g., Zolo-
1345 tov and Shock, 2001; Lowell and DuBose, 2005; Vance et al, 2007, 2016) and
1346 oxidized surface (e.g., McCord et al, 2002; Paranicas et al, 2009) may provide a
1347 source of redox energy for a subsurface biosphere. Life on Earth may have begun in
1348 relatively anoxic conditions potentially similar to conditions on Europa (Barge and
1349 White, 2017; Russell et al, 2017). The surface of Europa is littered with spectrally
1350 detected salts that include $MgSO_4$, $NaCl$, and SO compounds as well as sulfur
1351 implanted from the Io torus (see Carlson et al (2009) for review), and CO and
1352 CO_2 have been observed on the other moons but are likely unstable on Europa's
1353 surface (Hibbitts et al, 2000, 2003). Whether the ocean is highly reduced through
1354 interactions between the ocean and the seafloor (e.g., Zolotov and Shock, 2001;
1355 Vance and Goodman, 2009; Vance et al, 2014) or acidic due to downward trans-
1356 mission of oxidants from the surface (Pasek and Greenberg, 2012) depends heavily
1357 on surface geology and the exchange rate with the ocean. However, it is unclear
1358 how much of the surface is actually drawn down into the interior and how efficient
1359 surface-subsurface mixing could be. Determining what fraction of the surface is
1360 recycled or reprocessed in place, and how, will be important to constrain with
1361 *Europa Clipper*, alongside the elemental composition of surface materials. Overall,
1362 conditions of Europa's formation and later bombardment, as well as its past and
1363 potentially present activity, suggest the ingredients for life and present day energy
1364 to support it may exist.

1365 For Enceladus, direct measurements of its ocean composition have been made
1366 by the *Cassini* mission. The ocean likely contains compounds that suggest on-
1367 going interaction between the seafloor and ocean, as well as potential fuel for
1368 chemosynthetic life. Ocean-derived material erupting from Enceladus' south pole
1369 contains both simple and complex organics, ice crystals, and salts (Waite et al,
1370 2006, 2009, 2017; Postberg et al, 2011, 2018b), while Hsu et al (2015) demon-
1371 strated the presence of silica nanoparticles that are interpreted as evidence for extensive
1372 hydrothermalism and a well-mixed ocean. Waite et al (2017) demonstrated the
1373 presence of molecular hydrogen that would be available energy for metabolism;
1374 however, it bears questions as to whether hydrogen would be detected if the ocean
1375 were inhabited since this material could be consumed by life. Nonetheless, Ence-
1376 ladus' ocean geochemistry is the only measured to date and contains products

1377 known to support life on Earth, and therefore is habitable by current standards.
1378 A mission that would return to Enceladus and search for higher chained organics
1379 and organic complexity *in situ* within the plume could reveal strong indications
1380 as to whether life is present on Enceladus (e.g., Lunine et al, 2018).

1381 For Ganymede and Titan, where high pressure ice phases become important,
1382 the most pertinent question is whether any interaction can be maintained between
1383 the silicate layer and the oceans. Though Ganymede's outer shell is presently
1384 inactive at the surface, its deeper ice layers may overturn under basal heating
1385 from the deeper interior (e.g., Kalousová et al, 2018). Vance et al (2014) showed
1386 that the phase behavior of water-MgSO₄ salt mixtures under Ganymede conditions
1387 can form multiple ocean layers separated by high pressure ice layers, potentially
1388 with a deep reservoir of saline liquid above the silicate core that may argue for
1389 the possibility of serpentinization. Similar conditions could be possible at Titan,
1390 where clathrates may also play a role (Castillo-Rogez and Lunine, 2010) and the
1391 ice shell may not be convecting (e.g., Nimmo and Bills, 2010). It is unclear whether
1392 sources of oxidants could exist within ocean planets with outer ice shells in the
1393 stagnant lid regime, like Ganymede and Titan. However, if mixing between these
1394 reservoirs is possible, or if the decomposition of clathrates can deliver new sources
1395 of energy, these moons could be habitable. Titan's surface habitability is a different
1396 question all together; with liquid ethane and methane across the surface, Titan
1397 could support exotic kinds of habitability, fueled by different chemical compounds,
1398 including benzene (see Lunine et al (2019) for a review).

1399 6.2 *In Situ* Exploration of Terrestrial Habitats

1400 Ice on Earth is rich with life both within the ice and along its interfaces (e.g., Priscu
1401 and Christner, 2004; Deming and Eicken, 2007). Analog habitats in Antarctica in-
1402 clude perennially ice-covered lakes (e.g., Priscu et al, 1998; Murray et al, 2012) and
1403 subglacial lakes such as Vostok and Whillians (Christner et al, 2014; Mikucki et al,
1404 2016). Although communities are supported primarily by the silicate materials at
1405 the bed, most of the communities are chemosynthetic, or rely on the oxidation and
1406 reduction of species such as iron, sulfur, nitrogen and methane as metabolisms. In
1407 the Arctic, subglacial volcanos in Iceland (e.g., Gaidos et al, 2004) and sulfur-rich
1408 subglacial springs expressed at the surface of Borup Fjord (Gleeson et al, 2011,
1409 2012) power similar communities.

1410 A growing amount of research is conducted via field studies with analog mis-
1411 sion technologies, from *in situ* and remote sensing instrumentation to full vehicle
1412 platforms. Drilling vehicles are being developed for future missions as well as deep
1413 glacial access. In addition to ocean gliders (e.g., Lee and Rudnick, 2018) and open
1414 ocean AUVs (e.g., Jenkins, 2010; Dutrieux et al, 2014) that have been devel-
1415 oped for oceanographic operations on Earth, vehicles developed with ocean world
1416 exploration beyond Earth in mind include BRUIE (Berisford et al, 2013), EN-
1417 DURANCE (Gulati et al, 2010), ARTEMIS (Kimball et al, 2018), Nereid Under
1418 Ice (German and Boetius, 2019), and Icefin (Meister et al, 2018) and have operated
1419 under either lake ice, sea ice and/or ice shelves. These vehicles are returning data
1420 about physical and oceanographic processes as well as communities under the ice,
1421 including ice-ocean exchange.

1422 6.3 Future Exploration

1423 Future missions can investigate ice-ocean exchange in the outer solar system.
1424 NASA’s planned *Europa Clipper* mission concept would conduct multiple science
1425 investigations (Buffington et al, 2017). *In situ* sampling of Europa’s atmosphere
1426 would look for compositional and isotopic signatures of ocean pH and water-rock
1427 interactions. Remote sensing from the ultraviolet to mid-infrared would provide
1428 global mapping of 95% of the surface to better than 50 m/pixel. Ice-penetrating
1429 radar would sound through kilometers of ice to characterize ice shell structure and
1430 search for the ice-ocean interface. Gravity science, magnetometer/plasma, and
1431 altimetry investigations to constrain the thickness of the ocean and underlying
1432 silicate layer, as well as the vigor of tidal heating, are also important for under-
1433 standing the workings of the ice relative to the underlying materials (Pauer et al,
1434 2010; Verma and Margot, 2018; Steinbrügge et al, 2018). A complementary suite
1435 of instrument investigations are planned to fly on ESA’s *JUICE* spacecraft Grasset
1436 et al (2013), which is set to orbit Ganymede in the early 2030’s.

1437 Landed geophysical investigations on Europa (Pappalardo et al, 2013; Hand
1438 et al, 2017) could more precisely evaluate the satellite’s radial density structure,
1439 rheological state and thermal state of its mantle, and extent of seafloor hydration
1440 (Vance et al, 2018a,b). At the time of this writing, the *InSight* mission is con-
1441 ducting a geophysical investigation of Mars, using seismometers and heat probes
1442 to further characterize the radial structure, composition, and temperature of the
1443 interior (Smrekar et al, 2018). The *DragonFly* concept, selected as NASA’s next
1444 New Frontiers class mission, would include seismic and mass spectrometer inves-
1445 tigation that could reveal the hydration state of silicates in Titan’s interior and
1446 the degree of ice-ocean exchange (Turtle et al, 2017).

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